SEDIMENTATION OF THE MIDDLE PRECAMBRIAN

TYLER FORMATION OF NORTHCENTRAL WISCONSIN

AND NORTHWESTERN MICHIGAN

A THESIS

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ABSTRACT

The Tyler Formation crops out in a northeasterly-trending belt in Northcentral Wisconsin and Northwestern Michigan along the Gogebic Iron Range. Good exposures are found along several of the major streams draining the area, road cuts and railroad right-of-ways.

The Tyler is considered Late Middle Precambrian in age, the Wisconsin equivalent of the Baraga Group of Michigan's Marquette Range Supergroup. The formation was warped and slightly metamorphosed during the Penokean Orogeny (1.7 b.y.). Primary sedimentary structures have been generally preserved.

The sedimentologic aspects of seven outcrops in the Hurley, Wisconsin area were studied in detail and a measured stratigraphic section was established. Three distinct lithologies are present, sandstone (graywacke), siltstone and shale in varying degrees of metamorphism. Forty-one percent of the beds measured are argillites or slates and fifty-nine percent are sandstones or siltstones. Volumetrically, the sandstones and siltstones are more abundant than the shale units. A thinning-upward trend in the shale beds suggests a decreasing rate of sedimentation up section which in turn suggests increasing tectonic stability in the source area relative to the depocenter. Alternatively, the thinning-upward trend in the shales may reflect increased periodicity of the events which deposited the coarser-grained clastic beds. A corresponding volumetric increase in sandstones and siltstones up section is explained by the

peculiarities of local basin bottom topography.

Primary sedimentary structures including alternation of sand-mud units, laterally extensive bedding, graded bedding, microcross-bedding, sole marks, rip-up clasts and Bouma sequences suggest a turbidity current mechanism for sediment transport and deposition in a realtively deep water environment. Twenty-four percent of the beds studied in detail are recognizable turbidites while thirty-five percent do not contain specific telltale sedimentary structures. A grain-flow origin for at least some of these latter beds is suspected.

Because of the similarity between the lithologies and facies of the Tyler Formation and the ideal facies sequences of submarine fans, it is suggested that that part of the Tyler Formation which was studied in detail was deposited as part of a submarine fan complex.

Indicators of current movement found in the rocks of the Tyler are of several types including both interstratal and intrastratal. Sole marks and cross-bedding are most useful. The currents which deposited the Tyler sediments moved from the east-southeast toward the west-northwest. A new type of sole mark, called a ridge mold or negative groove, is described and a possible method of formation is suggested.

Two different methods for estimating current velocity were employed, one based on maximum clast size and one based on spacing of ridge molds. Both methods yielded velocities of a few tens of centimeters per second and primary slope is estimated at $0^{\circ}10'$.

Petrology reveals that the major framework constituents of the Tyler graywackes are quartz, plagioclase and rock fragments set in a chlorite - and mica-rich matrix. The "average" graywacke is a lithic graywacke with 28% matrix. Quartz and chert comprise 73% of the framework grains, rock fragments 17% and feldspar 10%. The source terrain was probably mostly granitic with some contribution from older sedimentary, metamorphic and volcanic rocks. The Lower Precambrian rocks to the south and southeast of the Tyler outcrop belt were the probable source.

Paleoslope was probably at right angles to current flow. Thus a southern limit to the extent of the Middle Precambrian depositional basin is defined. Similarity of the Tyler and other Middle Precambrian sedimentary rocks in the Lake Superior region suggests a common depocenter in a cratonic basin but at different depositional loci. Reconstruction of current movement and probable source area for the Tyler, Rove and Virginia (Thomson) Formations suggest that the depocenter was landlocked on three sides but may have been open to the northeast.

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RAPIDS ON THE BAD RIVER NEAR PENOKEE GAP,
MELLEN, WISCONSIN.
Water falls over Tyler Formation and small igneous intrusion.
(NE 1/4, SW 1/4, S11, T44N, R3W)

INTRODUCTION

The Tyler Formation is a thick sequence of interbedded slates and graywackes of Middle Precambrian age. The United States Geological Survey considers the Tyler a part of the Baraga Group of the Marquette Range Supergroup (Cannon, personal communication, 1976). The rocks of this formation crop out in a northeasterly trending belt approximately 50 miles long and one to two miles wide from the vicinity of Mineral Lake, eight miles west of Mellen, Wisconsin to just east of Ramsay, Michigan (Fig. 1).

The sedimentary aspects of these rocks were studied in the field and in the laboratory to gain some understanding of the possible source areas of the sediments, the sense of the paleocurrents which deposited them, transport mechanisms, and the environment of deposition. It is hoped that this study will prove useful to the geologists confronting the still unanswered questions of correlation, environments of deposition and basin configuration in the Middle Precambrian of the Lake Superior area.

I wish to gratefully acknowledge the patient professional guidance of Dr. Richard W. Ojakangas of the University of Minnesota - Duluth who brought the problems of Middle Precambrian geology to my attention, suggested this specific study and served as my advisor for the entire project.

I wish also to thank the Wisconsin Geological and Natural History Survey for providing the topographic maps necessary for my field work

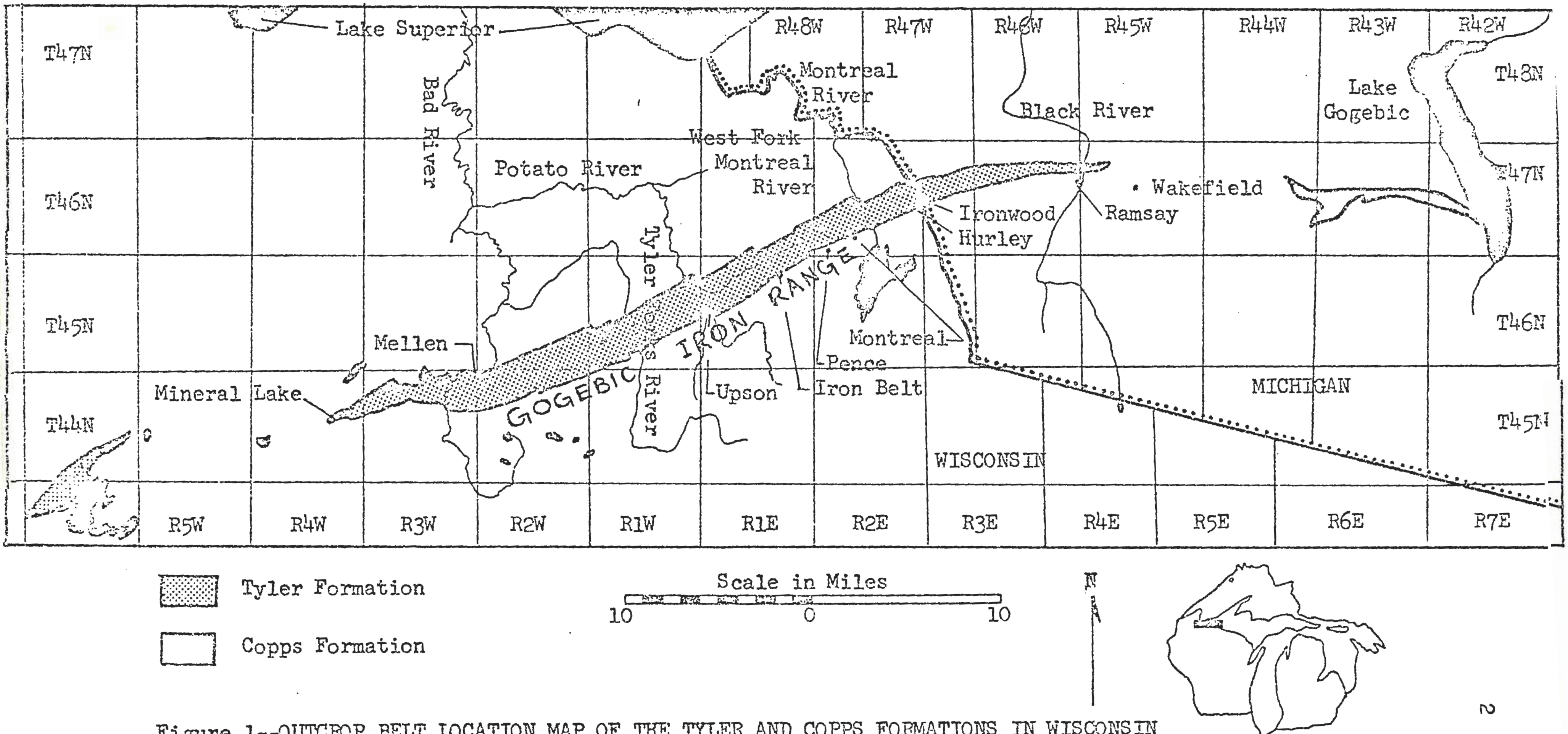

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Figure 1--OUTCROP BELT LOCATION MAP OF THE TYLER AND COPPS FORMATIONS IN WISCONSIN AND MICHIGAN. Geology from Dutton and Bradley, 1970; Bodwell, 1972; Martin, 1936.

and most of the thin sections used. Discussions with Dr. Ralph W. Marsden of U. M. D. were very beneficial. Dr. Marsden and Dr. G. B. Morey of the Minnesota Geological Survey were also members of my advisory committee, read the manuscript and made valuable suggestions.

My thanks are also offered to the entire faculty of the geology department at U. M. D. for their professional assistance and to my fellow graduate students for many helpful suggestions. The University of Wisconsin Geology Museum under the direction of Dr. Klaus Westphal loaned thin sections from their collection. The geology department of U. M. D. provided a few thin sections.

My deep appreciation and sincere thanks are extended to my fiancee, Ms. Lola Stenseth of Madison, Wisconsin for typing drafts of this paper and for her unfailing support throughout the period when this paper was in preparation.

GEOGRAPHY AND PHYSIOGRAPHY

The Tyler outcrop belt lies coincident with and just north of the Penokee-Gogebic Iron Range (usually referred to as the Gogebic Iron Range) of the Lake Superior iron and copper mining district. The range and the Tyler lie partly in Wisconsin and partly in Michigan with fully 2/3 of the belt on the western side of the Montreal River which forms the Wisconsin-Michigan boundary in the area. The Iron Range itself trends northeasterly, is approximately 80 miles long and lies 25 to 30 miles south of the southern shore of Lake Superior. The towns of Mellen, Upson, Iron Belt, Pence, Montreal and Hurley in Wisconsin, and Ironwood, Bessemer, Ramsay and Wakefield in Michigan are the principle population centers along the range and owe their existence to iron mining which began in the early 1880's. All of these towns are built either on the Ironwood Iron Formation or on the overlying Tyler Formation.

The topography in the approximately one hundred square miles covered by this report is a direct reflection of the structure of the rocks underlying the area, with but few anomalies caused by glacial deposits. The Tyler Formation is one of the rocks on the Iron Range most susceptible to weathering and erosion. The surface expression of this formation is therefore a valley trending parallel to the strike of the formation (northeast-southwest) and bounded on both the south and the north by long parallel ridges cored by more resistant rocks, the Palms and Ironwood Formations to the south and the Keweenawan lava flows to the north. These ridges are not continuous but are serrated by the major rivers which occupy fault zones crosscutting the Middle Precambrian rocks. Relief between the ridges and the Tyler valley is everywhere on the order of 100-200 feet. One of the highest hills in Wisconsin, Mount Whittlesey (1872 feet above sea level) forms part of the southern valley wall near Mellen, Wisconsin. The northern wall of the valley is generally steeper than the southern.

The divide between the Mississippi and St. Lawrence drainage basins lies just south of the Gogebic Range in Early Precambrian terrain. The waters of the major streams draining the study area, the Bad, Tyler Forks, Potato, and Montreal Rivers, flow northward into Lake Superior and from there to the St. Lawrence River. Since the northward course of the streams is almost normal to the structure and since the courses of the smaller and less competent tributaries are parallel to and a direct result of the structure, a crude trellis-type drainage pattern is established.

The only good natural exposures of the Tyler Formation occur along the major streams. Nowhere is a complete section exposed. The Tyler was named for the numerous outcrops along the banks of the Tyler Forks River (Van Hise, 1901). These outcrops are, however, highly weathered and so covered with undergrowth and lichen that the detailed examination of internal sedimentary features is very difficult. Of more use are the man-made outcrops along railroads and highways. A number of this latter type are to be found near Hurley, Wisconsin. The excellent exposures there have facilitated the detailed description necessary to establish the type section of the Tyler.

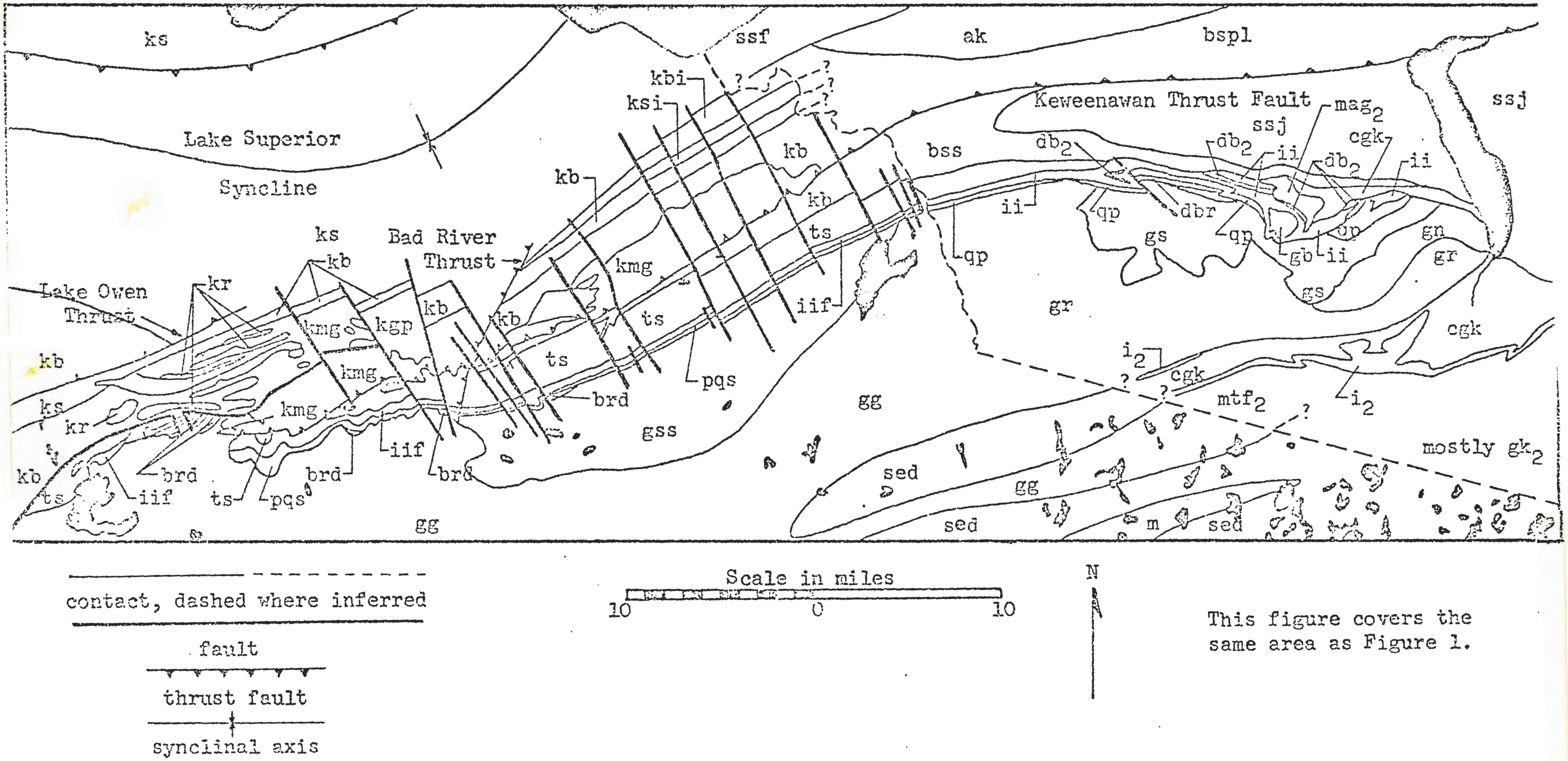
GEOLOGIC SETTING

The Precambrian Shield of North America extends southward from Canada into Minnesota, Wisconsin and Michigan. In northern Wisconsin and the upper peninsula of Michigan the shield is characterized by metamorphosed sedimentary and volcanic rocks and by granitic crystalline intrusive rocks (and local mafic intrusive rocks) of Early Precambrian age (Precambrian W using U.S.G.S. terminology), sedimentary and volcanic sequences of Middle Precambrian age (Precambrian X), and very thick piles of Late Precambrian age (Precambrian Y) extrusive rocks and overlying sandstones. Traveling from south to north through Ashland and Iron Counties of Wisconsin and Gogebic County of Michigan (which include all of the present study area) one crosses successively younger Precambrian rocks. The Upper and Middle Precambrian rocks strike northeast-southwest and dip steeply northward. On a regional scale the rocks form a structural homocline dipping northward toward Lake Superior and, as such, are part of the southern limb of the Lake Superior syncline (see Fig. 2).

In Wisconsin the Lower Precambrian rocks consist of heretofore undifferentiated granites, granitic gneisses and greenstones, the interrelationships of which are very complex. Recent mapping in Michigan by the U.S.G.S. (Schmidt, 1972, and others) has recognized the presence of at least four different ages within the Lower Precambrian crystalline rocks. The oldest rocks are the greenstones which have been intruded

EXPLANATION

		Wisconsin		Michigan	
Cambrian (?)				ssj	Jacobsville sandstone
	Late Keweenawan	ks	sedimentary rocks	ssf	Freda sandstone
	Middle and late Keweenawan	kbi, ksi	interbedded sedimentary rocks (ksi) and lava flows (kbi)	ak	andesite
		kgb	granite porphyry (Mellen Granite)		
Late Precambrian	Middle Keweenawan	kr	granophyre		
		kmg	Mellen Gabbro		
		kb	mafic extrusives	bpsl	Portage Lake Lava Series
				bss	South Range Lava Series
	Early Keweenawan		sedimentary rocks (not shown) including Bessemer Formation		
			unconformity		
	Baraga Group	ts	Tyler Formation	ts	Tyler Formation
				cgk	Copps Formation
Middle Precambrian	Menominee Group	iif	Ironwood Iron Formation	ii	Ironwood Iron Fm.
		pqs	Palms Formation	qp	Palms Formation
			unconformity		
	Chocolay Group	brd	Bad River Formation	dbr	Bad River Formation
			unconformity		
Early Precambrian		gg	granite and gneiss	gr	granite
				gn	granitic and dioritic gneiss
		gss	greenstone and greenstone schists	gs	greenstone
				gb	volcanics
Age and correlation uncertain		sed	undifferentiated & metamorphosed sedimentary rocks	db_2	diabase
				i_2	iron formation
		m	metamorphosed mafic intrusive and extrusive rocks	gk_2	graywacke
				mtf_2	mafic tuff
				mag_2	mafic agglomerate (Presque Isle Volcanics)



This figure covers the same area as Figure 1.

Figure 2--REGIONAL GEOLOGIC MAP OF NORTHCENTRAL WISCONSIN AND NORTHWESTERN MICHIGAN. Geology from Dutton and Bradley, 1970; Bodwell, 1972; Martin, 1936.

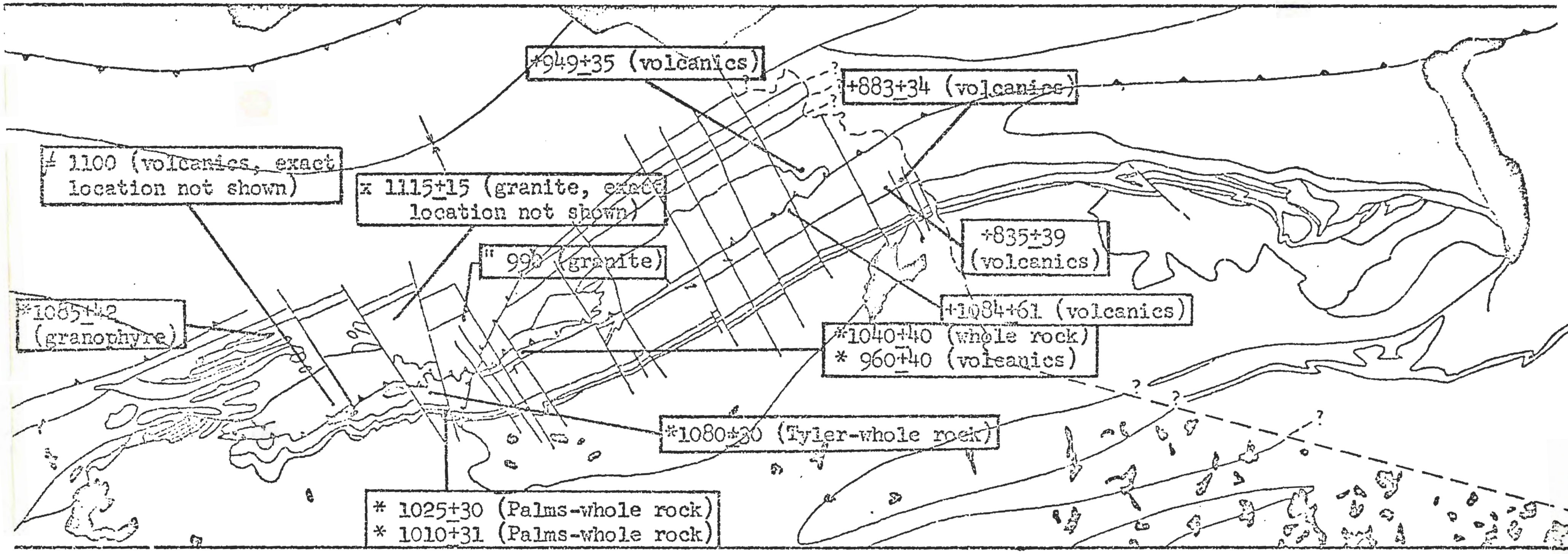
by several generations of quartz monzonites and granitic gneisses.

The sequence and lithologies of the Middle Precambrian sedimentary rocks which unconformably overlie the Lower Precambrian rocks in the Mellen-Hurley-Ironwood area resemble the Middle Precambrian sequence of Michigan's Upper Peninsula. The Michigan rocks belong to the Marquette Range Supergroup composed of the Chocolay, Menominee, Baraga, and Paint River Groups in order of decreasing age (Cannon & Gair, 1970). It is generally accepted that the Middle Precambrian rocks of the Gogebic Range are part of a similar but incompletely preserved sequence and tentative correlation with the Marquette Range Supergroup has been proposed (Table 1). In a like manner the Animikie Group of Minnesota and Ontario is often referred to as an equivalent series. In the absence of fossil evidence the geologist has had to rely upon gross lithologic and sequential similarities for correlation of these Middle Precambrian rocks. More recently, radiometric techniques have provided reliable dates for extrusive and intrusive rocks which cut the Middle Precambrian sedimentary rocks and for the metamorphism of those sedimentary rocks (see Fig. 3).

Young (1966) and Church and Young (1972) propose an overlap relationship between the Animikie of Minnesota and the Wisconsin-Michigan sequences (previously called Huronian), with the Lower Animikie being roughly equivalent to the Upper Huronian. Morey (1973) states that the Animikie of Minnesota is similar to the Penokee (Gogebic) sequence in Wisconsin and both are dissimilar to the type Huronian

TABLE 1--GENERAL STRATIGRAPHY AND CORRELATION CHART FOR THE PENOKEE-GOGEBIC, MARQUETTE, AND MENOMINEE IRON RANGES, WISCONSIN AND MICHIGAN

ERA	PERIOD	OROGENY	ABSOLUTE AGE (b.y.)	PENOKEE-GOGEBIC RANGE	MARQUETTE RANGE	MENOMINEE RANGE		
				sandstone	sandstone			
		Grenville	1.1	gabbro & granite		rare diabase dikes		
Late Prec.	Keweenawan			sandstone, shale & conglomerate				
				volcanics				
				sandstone (Bessemer)				
		Penokian	1.7	granite	granite & basic intrusives	granite & gabbro		
Middle Precambrian	Huronian (?)					Fortune Lakes Slate Stambaugh Fm. Hiawatha Graywacke Riverton Iron Fm. Dunn Creek Slate	Paint River Group	Marquette Range Supergroup
				Tyler Formation (12,000'+)	Michigamme Slate Clarksburg Volcs. Greenwood Iron Fm. Goodrich Quartzite	Badwater Grnste. Michigamme Slate Hemlock Volcs. Michigamme Slate	Baraga Group	
				Ironwood Iron Fm. (850') Palms Fm. (300'-800')	Negaunee Iron Fm. Siamo Slate Ajibik Quartzite	Vulcan Iron Fm. Felch Fm.	Menominee Group	
				Bad River Formation (0-300', 400' or 550') Sunday Formation (0-150')	Wewe Slate Kona Dolomite Mesnard Quartzite	Randville Dolo. Sturgeon Qtzte.	Chocolay Group	
		Algoman	2.5	granite (2.5?)	gneiss (?)	gneissic granite (?)		
Early Prec.		Laurentian	(?)	granite (?)	gneiss (?)	gneissic granite (?)		
				greenstones	greenstones			



" Goldich, et al, 1961 (K-Ar)
x Silver & Green, 1963 (U-Pb)
≠ Chaudhuri, et al, 1969 (Rb-Sr)
* Komatar, 1972 (K-Ar)
+ Cooper, 1973 (K-Ar)

Ages in millions of years
Dates on Tyler & Palms Fms. interpreted as dates of metamorphism, Van Schmus, et al. (1975) shows Tyler as 1900±50 m.y.

Figure 3--RADIOMETRIC AGE DATES ON SOME ROCKS OF NORTHCENTRAL WISCONSIN AND NORTHWESTERN MICHIGAN

(north shore of Lake Huron) in that the latter lacks significant iron formations. He, therefore, warns that the term Huronian be used with caution when referring to the Minnesota rocks. Cannon and Gair (1970) have proposed the name Marquette Range Supergroup for the Middle Precambrian sequence in Michigan's Upper Peninsula and for adjacent areas in Wisconsin. In this paper, the Tyler Formation of the Gogebic Iron Range will be considered part of the Baraga Group of the Marquette Range Supergroup (Table 1).

The oldest rock of Middle Precambrian age on the Gogebic Range is the Sunday Formation (quartzite) which overlies the crystalline Lower Precambrian rocks with angular unconformity (see Fig. 2). The Sunday Formation crops out only in Michigan where maximum thickness has been estimated at 150 feet (Van Hise & Leith, 1911; Schmidt & Hubbard, 1972). Neither this formation nor the conformably overlying Bad River Formation (dolomite) crop out continuously along the southern margin of the range. The Bad River Formation varies in thickness from a minimum of 0 to a maximum of 300 feet (Irving & Van Hise, 1892; Van Hise & Leith, 1911), 400 feet (Schmidt & Hubbard, 1972) or 550 feet (Komatar, 1972). A disconformity separates the dolomite from the Palms Formation (slate and quartzite), the next younger formation. The Palms exhibits rather continuous outcrop from one end of the range to the other and in many places forms the southward-facing cliffs and crests of the long serrated ridge which marks the Iron Range proper. The Sunday and Bad River Formations are missing locally, so the Palms is in contact with the Lower Precambrian rocks. Estimates of the average thickness of the Palms are 300-400 feet (Irving and Van Hise, 1892; Van Hise and Leith, 1911), 400-500 feet (Meek, 1935) and 400-450 feet (Schmidt and Hubbard, 1972). Maximum thickness is about 800 feet. Conformably overlying the Palms is the Ironwood Iron Formation which averages 850 feet thick (Van Hise and Leith, 1911; and others). In the eastern end of the district the Ironwood interfingers with the Presque Isle Volcanics which are 3500-4000 feet thick (Hendrix, 1960). It was the economic interest in the Ironwood Iron Formation which concentrated early geologic work in the Middle Precambrian sequence of this region.

The youngest sedimentary formation and by far the thickest of the Middle Precambrian sequence is the Tyler Formation. The nature of the contact between the Tyler and the Ironwood has been debated for some time. In Wisconsin the two formations look conformable except for the presence of a fragmental horizon at the base of the Tyler which, according to some, represents an erosional unconformity. Hotchkiss (1919) cites erosion as the sole cause of thickness variation in the Anvil Member (youngest) of the Ironwood Formation. The Anvil varies from 0 to 375 feet thick. The presence of an unconformity marked by a conglomerate (resting on volcanics) at the base of the Copps Formation (which occupies the same stratigraphic position as the Tyler just to the east of this study area, see Fig. 1 & 2) is uncontested although the lateral extent of this conglomerate is debatable. Atwater

(1935a, 1935b, & 1938) concludes that the Tyler and Copps Formations are correlative and suggests the presence of an unconformity between the Ironwood and Tyler on the basis of that found between the Ironwood (and interfingered volcanics) and the Copps. The Tyler-Ironwood contact is not exposed and I have seen nothing in the course of my work that I would call an unconformity although several fragmental horizons were located within the Tyler. A strong case for a gradational contact was made by Aldrich (1929). Not the least of the evidence he presents for support is the presence of iron carbonate far up into the Tyler (150 feet above the base). A conglomeratic zone like that in the Copps does not crop out in the Tyler area and, in fact, all references to the Pabst Conglomerate (the basal fragmental member of the Tyler) which were encountered in the reading were references to either drill core or to underground exposures in iron mines, neither of which I had access to. It is, furthermore, well known that conglomerates can change in compositional and textural character within very short distances along strike (e.g., Blatt, Middleton & Murray, 1972, page 197) let alone within the several miles between the easternmost exposure of the Tyler and the westernmost outcrop of the Copps (Fig. 1 & 2). However, because of documentation of a post-Ironwood, pre-Tyler period of tectonism by Hendrix (1960), Felmlee (1970) and Cannon (1973), in light of the facts cited above, and because I did not personally examine the Copps Formation, Atwater's correlation and the unconformable relationship between the Tyler and the Ironwood which it apparently necessitates are accepted for the purposes of this paper.

The Tyler is the youngest of the Penokee-Gogebic series and is overlain with angular unconformity (Schmidt & Hubbard, 1972) by the Lower Keweenawan Bessemer Formation (quartzite). It, in turn, is overlain by the very thick Middle Keweenawan volcanic pile which is succeeded by the Upper Keweenawan (or early Cambrian) sandstones of the south shore of Lake Superior.

It is impossible to estimate how much rock was eroded during the periods represented by the several unconformities cited. That some was removed is substantiated by the angular truncation of the Lower Precambrian structures, by the presence of jasper pebbles in the basal conglomerate of the Palms Formation where no jasper is known in the rocks immediately beneath it (Irving and Van Hise, 1892, page 455; and Van Hise and Leith, 1911, page 229), by the spotty nature of the occurrence of outcrops of the Sunday and Bad River Formations (suggesting erosional dissection of a once continuous outcrop belt, (Dutton & Bradley, 1970) and by identification of iron formation fragments (Huber, 1959) and milky quartz pebbles within the clastic Tyler sediments and the Lower Keweenawan conglomerate (Aldrich, 1929). The angular unconformity at the top of the Tyler and the regional beveling of the Tyler Formation at the eastern end of the exposure belt before deposition of the Keweenawan sedimentary and volcanic rocks (Fig. 2) attest to the folding and erosion which occurred during post-Tyler time. Thus, the Middle Precambrian rocks which remain may be only a portion of a once thicker sedimentary sequence.

The entire Middle Precambrian sequence strikes northeasterly and dips steeply (65-70°) to the northwest. Structural interruptions along strike include crosscutting faults and some large scale folding in T44N, R4 & 5W (Fig. 2). The faults are identifiable by magnetic survey of the Ironwood Formation (Aldrich, 1929) and in some cases, by offset of the lowermost Tyler Formation and the oldest Keweenawan flows (Fig. 2). The western terminus of the Tyler is marked by thrust faulting, although this has been recently disputed (see Felmlee, 1970), and intrusion of the Keweenawan Mellen Gabbro and granite porphyry. Because of the westward pinching of the Tyler Formation these Keweenawan rocks come into contact with the underlying Ironwood in the vicinity of Mineral Lake. Areas shown on the maps west of Mineral Lake as Tyler Formation are inferred from magnetic surveys (Aldrich, 1929).

In recent years students of Dr. C. Craddock at the University of Wisconsin-Madison have studied the structural intricacies of the Mellen area (especially Felmlee, 1970 and Cooper, 1973) and in general conclude that there is no evidence for the continuation into this region of the Keweenawan thrust fault (Fig. 2) known to exist both to the west and east. Rather, the contact between the Tyler and the granite and gabbro are believed to be of an intrusive nature. Emplacement of the gabbro complex was perhaps aided by the unconformity found between the Tyler and the oldest Keweenawan rocks (Tabet, 1974) and by the weakness of the sedimentary bedding planes within the Tyler. Hendrix (1960) suggested that the Keweenawan thrust fault passes through the Early Precambrian terrain south of the Middle Precambrian sequence.

To the east, the Tyler belt pinches out as a result of erosion which, subsequent to broad regional warping and elevation, completely removed the Tyler in that vicinity. Here the overlying Keweenawan sedimentary and volcanic rocks are in contact with the underlying Ironwood Formation just east of Ramsay, Michigan (Fig. 2).

Diabase dikes and sills, granophyre and small outlying plutonic bodies intrude the entire Middle Precambrian sequence, the lowermost Keweenawan flows and the granite and gabbro. Most are probably of Keweenawan age but some are truncated by the Keweenawan flows indicating a Late Middle Precambrian time of emplacement.

Several authors have suggested the correlation of the Tyler Formation with the Rove, Thomson, Virginia and Rabbit Lake Formations of Minnesota since the original suggestion by Irving (1880). These formations occupy similar stratigraphic positions in Middle Precambrian rock sequences of slates, quartzites and iron formations. Lithologically, they are very similar. The present dogma among Lake Superior geologists is to consider iron formations correlative, a doctrine which would suggest that the overlying Rove, Virginia, Rabbit Lake and Tyler Formations are correlative units as well. Further, it has been suggested that these formations all occupied different loci in a common depositional basin.

Goldich, Nier, Baadsgaard, Hoffman and Krueger (1961, page 118) suggested the existence in Middle Precambrian time of a large geosynclinal trough extending from Minnesota through Wisconsin to Michigan and possibly connected to the Labrador trough through the Mistassini district. The Animikie rocks of the Mesabi Range represent the miogeosynclinal deposits of this basin. These sediments and later intrusive bodies became the framework for a belt of highlands called the Penokean Mountains which were formed during the Late Middle Precambrian Penokean Orogeny (1.7 b.y.) and were coincident with the ancient geosyncline. This hypothesis would suggest that the Tyler Formation, by reason of geographic location, was part of the eugeosyncline. It is the author's opinion that the Tyler Formation is not eugeosynclinal in character.

STUDY METHODS

Field Work

Precursory examination of the Tyler Formation by Dr. Richard Ojakangas and by the author led to the not unwarranted assumption that the Tyler resembles a class of sediments called turbidites. Because of this preconceived notion, preparation for the field work was done with the idea of recording the descriptive details of the myriad of sedimentary features unique to this class of sediments as well as the normally recorded characteristics of other sedimentary rocks.

The field work began in the early fall of 1973 and continued to include the entire field season of 1974. Likely outcrop locations were first pinpointed by reference to previous publications (especially Aldrich, 1929) and by identification of areas on topographic maps exhibiting steep slopes and substantial relief. The method of outcrop location was then to drive every passable road, walk old logging trails and the banks of all the major streams and traverse the Tyler valley in such a manner as to examine the probable outcrop locations identified from topographic maps and other publications. Over 80 outcrops were found and examined; the best outcrops were measured and described in detail.

Each outcrop description included the following: location, physical dimensions and appearance (with notes as to the ease of seeing detailed sedimentary features), a short lithologic description including composition, size, roundness, and sorting of grains, color, thickness of the sedimentation unit, bedding attitudes, slaty cleavage attitudes and the presence or absence of other structural features like joint systems and small scale folding and faulting.

If the outcrop proved to be of such quality as to make the observation of internal sedimentary structures easy (and in general these outcrops were only the man made ones) a data sheet (Appendix A) was filled out for each bed, including the stratigraphic number assigned to the bed, the presence or absence of the internal units of the Bouma classification (Bouma, 1962), the thicknesses of the Bouma units and the nature of the contacts between them (whether sharp or gradational), and the orientations of various paleocurrent structures. Included were cross-bedding, ripplemarks, sole marks, flame structures, parting lineations and other miscellaneous features. Convoluted beds were examined for orientation of convolute axes. Various types of graded bedding were recognized.

Special attention was given to several outcrops in the vicinity of Hurley, Wisconsin in an effort to piece together and describe the most complete stratigraphic section possible. That section has been drawn to scale and is presented later.

An effort has been made throughout the sampling which accompanied this study to be cognizant of the difference between the target population (the entire Tyler Formation) and the sampled population (that part of the Tyler which is exposed). The scattered nature of the outcrops examined requires that inferences drawn from the conclusions of this work apply only to the sampled population. Some thought was given to the possibility that the outcrops found are atypical and that most of the Tyler, being buried beneath glacial ground moraine, is of a different lithology than that part exposed. Aldrich (1929) estimated that the ratio of total area of exposure to total area of his survey (the entire Middle Precambrian sequence) was on the order of 1:100,000. I consider the exposures studied to be representative of the entire Tyler Formation for these reasons: the manmade outcrops clearly show that even the more resistant graywacke beds (as opposed to the slate beds) do not crop out naturally; the plutonic intrusive rocks at the west end of the belt have metamorphosed the Tyler without regard to lithologic type and the induration thus imparted to the Tyler has resulted in a fairly representative outcrop sampling of all lithologies within the formation.

It was imperative for statistical reasons that samples be collected in a random nature, yet not without some attention to assure that each distinct lithology is represented in the final summary. At least one sample was taken from each outcrop to achieve lateral randomness. At those outcrops where detailed studies were made some bias was introduced. While attempting to retain randomness vertically through the measured section, special attention was given to sampling the coarser sediments (since slates under the petrographic microscope yield little valuable information) and to sampling those beds which contained paleocurrent indicators. Comparisons of the compositions of these latter beds with their paleocurrent directions should prove especially useful in the reconstruction of the possible source areas of the sediments.

Lab Work

In the lab 169 hand samples and thin section heels were stained for potassium feldspar and the estimated percentages recorded. One hundred thin sections of samples collected by the author and by others (see Appendix B) were examined and modal analyses were determined for 27.

From the information recorded on the field worksheets and the field notebook a straigraphic section and other figures were drawn. Special attention was given to the structure in the Hurley area to enable the author to piece together the several columns measured in the field.

PREVIOUS WORK

With the exception of the work by Atwater on the correlation of the Tyler and Copps Formations (Atwater, 1935a) and the early U.S.G.S. work, all work to date on the Tyler has been the result of interest in the geologic and economic character of the adjoining formations or in the regional structure of the southern shore of Lake Superior. No one has made a comprehensive description of the sedimentation, paleocurrents, source area and environment of deposition of the Tyler Formation. A type section has never been established. The publications cited below should not be construed to be a comprehensive list of publications on the Tyler Formation. Frequent reference to the maps in this paper will be beneficial during the following discussions.

Early Exploration, 1840-1884

The earliest recorded geological work in the Tyler valley was done by Houghton, commissioner of the state of Michigan, in the company of Captain Crane of the United States topographical engineers while on a survey of the Montreal and Menominee Rivers in 1840. These men undoubtedly discovered the outcrops of the Tyler on the Montreal River but the references available did not substantiate that fact.

In 1845 and 1846, C. Whittlesey of Cleveland, Ohio, in possession of Houghton's field notes, explored the area farther to the west along the range from the Montreal River to the Bad River. A. Randall in 1848 discovered iron formation on the range while accompanying the party of men engaged in the survey extension of the 4th Principle Meridian to the shore of Lake Superior. He took a specimen of magnetic iron formation from Township 47 in Michigan. Whittlesey returned in 1849 as a member of D. D. Owen's geological exploration party in Wisconsin, Iowa and Minnesota under the auspices of the United States Geological Survey. On this trip Whittlesey explored the western branches of the Bad River as far west as Lac des Anglais (today English Lake). A report of his work was published in 1852 as a part of the geological report by Owen (Whittlesey, 1852; Owen, 1852).

The spring of 1859 saw a visit to the range area by Daniels of the Wisconsin Geological Survey and I. Lapham representing the Milwaukee Iron Company. The purpose was to examine the economic potential of the iron formation and to begin a survey for the construction of a railroad to the range. This exploration covered the area from Tyler Forks River to Section 24, T44N, R4W in Wisconsin. Lapham reported a lessening of dip in the Middle Precambrian rocks west of Bad River. R. D. Irving (Irving and Van Hise, 1892) gives credit to Lapham for first discovering the fault at Penokee Gap (a natural breach in the Iron Range through which the Bad River flows in section 14, T44N, R3W). It was through this gap that the Wisconsin Central Railroad, today the St. Paul and Sault Ste. Marie, was built in 1873. Rail service branched to the center of the range in 1887, three years after ore shipments began. Lapham made the first map of the Penokee series.

However, the emphasis of that map was on location of the iron ore and the Tyler Formation was largely ignored.

In 1860 Whittlesey returned to the region as a representative of the Wisconsin Survey and although the survey never published his report, his findings, accompanied by a map, were eventually released in the Proceedings of the Boston Society of Natural History (Whittlesey, 1863). On this map and in the text of his report Whittlesey described Formation 4 as a silicious unit composed of two members. Both members were quartz-rich, slaty and in layers or beds. The lower unit was, however, dark colored and separated from the upper by a "bed" of magnetic iron and iron slate. The upper unit was lighter in color, novaculitic in places and had a N60°E to N65°E strike and variable dip (from 30° to 75°) to the northwest. The lower unit is today known as the Palms Formation and is separated from the Tyler Formation (the upper unit) by the Ironwood Iron Formation.

It was Whittlesey who first called the range the "Pewabik Iron Range" a name which was incorrectly printed as "Penokee Iron Range" by the compositor. It was also Whittlesey who named the main tributary to the Bad River the Tyler Forks, from which the Tyler Formation takes its name. Several of Whittlesey's publications on the range area are listed under REFERENCES CITED.

For the next few years investigations on the Penokee Range were undertaken by the Wisconsin Geological and Natural History Survey. E. T. Sweet and C. E. Wright made a reconnaissance of northern

Wisconsin under the direction of O. W. Wight, survey director in 1875. They crossed the range at Penokee Gap. T. C. Chamberlin took over the directorship in January of 1876 and instructed Wright and F. H. Brotherton to examine the iron series in detail between the Bad River and Lake Namekagon. It was during these years that R. D. Irving and C. R. Van Hise made frequent trips to the area. These trips were the field work upon which U.S.G.S. Monograph 19 was based (Monograph 19 is discussed in more detail later).

The first iron ore shipment was made from the Colby Mine (NE 1/4, S.16, T47N, R46W, Bessemer, Michigan) in 1884. By 1885 the Milwaukee, Lake Shore and Western Railway (later the Chicago and Northwestern) linked the range to the Ashland, Wisconsin ore docks.

Irving, 1877

R. D. Irving (1877a, page 23) wrote for the Annual Report of the Wisconsin Geological Survey:

> "The Penokee Iron Range has now been examined in greater detail than any other area of corresponding extent in the state... As far as purely geological work, without the aid of digging, is concerned, it may be safely said that the region is practically exhausted."

It was Irving's exploratory efforts on the range from the Potato River west to Penokee Gap and from there to Namekagon Lake (1873, 1876, 1877) and those of T. C. Chamberlin and A. D. Conover east of the Potato River to the Montreal River (1877), which prompted that statement. In the same report from which the above quotation is taken (Hotchkiss, 1878, page 25) one finds this reference to the Tyler:

> "...a belt of black slate lying north
> of the silicious schists..."

And, essentially, that is the extent of the description.

Conover returned to the area in 1878 to fill in gaps left by previous work, but he spent most of his time north of the Middle Precambrian series.

Irving and Van Hise, 1880

A more complete description of the Tyler is found in Irving's 1880 publication. That paper recognized several formations where today but one is recognized. Irving was the first to describe the spotted slates found in the railroad cuts at Penokee Gap and elsewhere (see PETROLOGY, this paper). He attributed the spots to remnants of crystalline chiastolite (andalusite). Irving also noted the incongruity of strike and dip of the Tyler from one end of the railroad cut to the other at Penokee Gap. This structural abnormality led to the mapping of a fault zone at that location. Offset along that fault was determined to be 800 feet in a horizontal sense and 1700 feet vertically with the west side up and the east side down. Regarding the structure of the entire Middle Precambrian sequence, he estimated total thickness at 13,000 feet and recorded the same lessening of dip to the west of Section 16, T44N, R3W as had Lapham before him. Irving (1880, page 103) hypothesized the following:

> "...we regard them [the Middle Precambrian rocks] as forming part of a great bend underneath the trough of Lake Superior and reappearing on the north side of the lake with a reverse inclination."

A. A. Julien did the petrographic work for The Geology of Wisconsin, 1873-1879, Volume III and reported the following mineral species in samples from Irving's Formations V, XII, XVI, and XXI (Julien, 1880): quartz, amphiboles, feldspars, biotite, tremolite, chiastolite, hematite, pyrite, magnetite, and limonite. Quartz, feldspar and biotite were most abundant.[1]

It fell to Wright (1880a) to summarize the Huronian west of Penokee Gap in the same publication. He attributed the pitting on the weathered surfaces of the spotted slates to the preferential weathering of micas in the rock. Wright (page 254) was the first to make the suggestion that several of Irving's numerous formations could be combined into one:

> "These [a group of micaceous quartz schists, micaceous slates and schists, and chiastolitic schist], with [Formation] No. XV, no doubt could consistently be comprised under one head..."

Irving and Van Hise, 1883[2]

T. C. Chamberlin (1883, page 83) offered in Volume I of The

[1]Hereafter, entire mineralogical descriptions of the Tyler will not be included with each author cited. Only mineral species not previously reported will be listed.

[2]Volume I of Geology of Wisconsin was published after Volume III.

Geology of Wisconsin, 1873-1879 this description of the Tyler Formation:

> "Upon the magnetic schists there repose a series of black, mica-bearing slates, alternating with diorites and schistose quartzites, including several horizons which are concealed by superficial material, and whose character is, therefore, unknown."

Chamberlin went a step farther than his predecessors when he drew the map of Huronian paleogeography reproduced in Figure 4 and he stated:

> "...they [the Tyler sediments] were unquestionably derived from the adjacent Laurentian [Late, Early Precambrian] land."[1]

This was the earliest attempt at reconstruction of the depositional history of the Tyler which I have found.

Irving and Van Hise, 1892

Monograph 19, published by the U.S.G.S., was the most authoritative work yet done on the Middle Precambrian series. This volume offered an historical summary of the previous geological work in the area, stratigraphic and lithologic descriptions of the formations, structural interpretations, petrologic details and some attempt at paleogeographic reconstruction and geologic history.

Besides the men already mentioned, Irving and Van Hise cited the following geologists who worked in Michigan:

[1]As quoted in Irving and Van Hise, 1892, page 64.

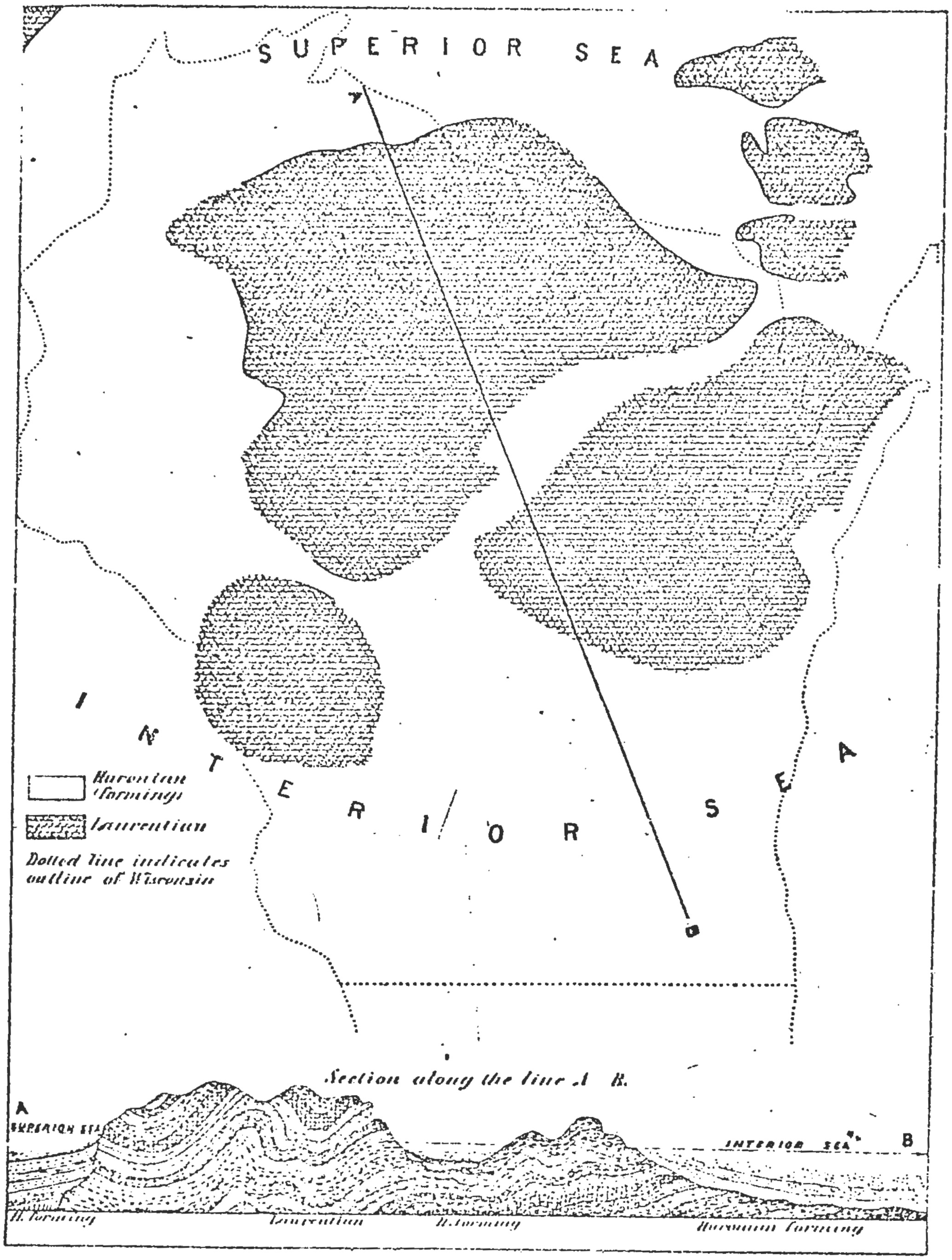


Figure 4 -- HYPOTHETICAL MAP OF THE LAND AREA IN THE HURONIAN (LATE MIDDLE PRECAMBRIAN) AGE.
From Chamberlin, 1883, page 80.

> G. O. Barnes, and Barnes and J. D. Whitney who explored the region between Gogebic Lake and the Montreal River in 1847 but mapped only volcanics and granite where the Middle Precambrian sediments lie (this error was reproduced on all maps showing the Michigan side of the river until 1872),
>
> T. B. Brooks and R. Pumpelly who covered a very large land area for the Michigan Geological Survey (from the Bad River in Wisconsin to the Ontonagon in Michigan) and whose efforts resulted in the first published map showing the Middle Precambrian in Michigan (Brooks, 1873),
>
> C. Rominger, Michigan State Geologist, who studied the region between the Montreal River and Lake Gogebic in 1882.

Monograph 19 grouped all the rocks above the Ironwood and below the volcanics into a single formation called the Upper Slate Member of the Penokee series. The Early Keweenawan quartzite (the Bessemer Formation) was not recognized as a separate formation. Quartzite outcrops were shown on the maps as part of the Upper Slate. The Upper Slate was described as mostly slate with minor quartzite and schist. Van Hise used the term graywacke on page 306 to characterize some of the rocks. The Ironwood-Upper Slate contact was described as gradational.

Structurally, the monograph recorded the same attitude for the Upper Slate Member as previous workers had, northeasterly strike with steep northerly dip. Thickness estimates ranged from 7150 feet at the Montreal River to 12,360 feet at the Tyler Forks River where the formation achieves its greatest exposed thickness.

Petrologic work reported only two minerals not listed by Julien 12 years before, muscovite and garnet. Four groupings of rock types were made including mica schists and slates, graywackes and graywacke slates, clay slates and phyllites, quartzite and conglomerate. The west end of the Upper Slate Member (west of Penokee Gap) had only mica schists and mica slates while the east end was characterized by micaceous graywackes and graywacke slates, chloritic graywackes and graywacke slates, and clay slates. The Upper Slate showed an increase in feldspar content to the west. Quartz overgrowths and secondary feldspars were recognized (suggesting metamorphism) but no rock fragments were reported. On the basis of the petrologic work it was suggested that the Southern Complex (the Lower Precambrian rocks to the south) was the sediment source.

With regard to the source area and mode of deposition of the graywackes and the slates, the authors made the point that since compositionally they were the same, since metamorphic changes in each were identical and since they were interstratified, their original environment of deposition (tectonic and surficial) must have been the same. No explanation for the striking grain size difference (clay versus sand and granule) in adjacent beds was offered.

Table 2 compares the geologic history of the Penokee-Gogebic range as published by Irving and Van Hise in Monograph 19 with that of several other authors.

TABLE 2--SUMMARY OF MIDDLE AND LATE PRECAMBRIAN HISTORY OF THE GOGEBIC RANGE AREA ACCORDING TO SEVERAL AUTHORS

Age	Irving & Van Hise 1892 (Wisc. & Mich.)	Van Hise & Leith 1911 (L.Superior area)	Aldrich 1929 (Wisc. only)		Atwater 1935 (Wisc. & Mich.)	Hendrix 1960 (East Gogebic)	Felmlee 1970 (Mellen area)	Cannon 1973 (Michigan)
Early Prec.	*							
Middle Precambrian		depo. of Sunday Fm. E.Prec. source rx						
	depo. of Bad River Fm. in quiet water		reports shallow water (algal) depo.					
	induration of Fms.							
	uplift & erosion removing most of Bad R. Fm. & possibly other Fms.							PENOKEAN OROGENY BEGINS; block faulting coincident with depo of Menominee Group
	depo. of Palms Fm. sed. source-Bad R. & Early Prec. rocks	also report jasper fragments in basal Palms	also reports graywacke fragments in basal Palms					
	rework upper Palms to clean qtz. sand	contemporaneous depo. of clastics & iron prec. with volcanism in east			Post Ironwood, Pre-Tyler orogeny with greatest uplift in East, less in West; depo. of Copps basal conglomerate grading into normal Tyler depo. to west unconformity	Post-Mid., Pre-Late Prec. epeirogeny characterized by block faulting initiated by subsidence	slight erosion	general uplift and erosion
	clastic sed. ceases chem. prec. begins depo. of Ironwood						Tyler depo. in deep water; if originally shallow then depo. & subsidence contemporaneously; if originally deep then little subsidence	depo. of Baraga & Paint River Groups in eugeosynclinal environment
	clastic sed. begins chem. prec. ceases depo. of Tyler Fm. sed. source-Early Precambrian rocks					subsidence cont'd. into Copps (Tyler) deposition		
	induration of Fms.		folding, faulting, no beveling (except in East); pinching of Tyler caused by faults; jointing; dikes injected into joints; all Mid. Prec. strata still flat-lying				PENOKEAN OROGENY folding & slaty cleavage in Tyler; compression from SW in east only low grade meta. (regional) in west; some dike & sill intrusions; broad arching in east, erosion of Tyler exposing Iron.	
	gentle folding, beveling of Tyler	folding, faulting beveling of Tyler				Post Copps (Tyler), Pre-Late Precambrian orogeny, axis of compression NW-SE; Sunday Lake Fault; intrusion of dikes, sills & bowing up of Mid.Prec. strata		uplift & deformation; END OF PENOKEAN OROGENY
Late Precambrian	depo. of Bessemer							
	tilting toward center of Lake Superior basin; pillow lavas subaerial lavas feeder pipes preserved as diabase dikes in Mid.Prec.rx.		lavas; downwarping into L.Sup. basin attendant longitudinal faulting (East) & folding (West) in Middle Precambrian				Mid. & Late Keweenawan igneous extrusives	
	Mellen Gabbro intruded, with many offshoots		thrust-faulting & warping of Mid. & Late Prec. due to torsional forces caused by tilting into L. Sup. accompanied by intrusion of Gabbro				probable cross faulting, intrusion of Mellen Gabbro	
		folding, faulting & peneplanation	faulting				fracturing & dikeing in Tyler; some isoclinal folding	
							granophyre intrusion	
			lavas					
	depo. of Late Prec. (Early Paleo. ?) sandstones						granite porphyry intrusion	
			cross faulting					

* a blank space indicates that the author suggested no change to that part of the summary of that his/her study did not deal with those events

Diller, 1898

In the text which accompanied the Educational Series of Rock Specimens was found a description of a typical graywacke sample from a railroad cut near Hurley, Wisconsin. A chemical analysis was given and the rock described as being composed of subrounded to subangular clastic particles of quartz, feldspars, hornblende, and rock fragments (in order of abundance) set in a matrix of fine quartz, biotite, magnetite, muscovite, kaolinite, and hematite.

Van Hise, 1901

Van Hise proposed the name Tyler Slate for the Upper Slate Member in 1901. No type section was established although the name was taken from the Tyler Forks River along which several exposures occur.

Van Hise and Leith, 1911

Van Hise and Leith followed in 1911 with publication of Monograph 52 entitled The Geology of the Lake Superior Region. This volume refers to the Tyler Formation as the Tyler Slate, a name which is still in use today (Keroher, et al, 1960, Lexicon of Geologic Names) and a name which mistakenly connotes a dominant lithology.

The names of Schoolcraft (who made visits to the range in 1821 and 1854), Norwood (1852 visit) and N. H. Winchell (Minnesota State Geologist in 1884) were added to the list of visitors to the south shore of Lake Superior. Winchell was one of the first (along with Irving, 1880a) to hypothesize the possible correlation of the Animikee Series of Minnesota with the Penokee Series of Wisconsin and Michigan.

Monograph 52 described the Tyler as dominantly a pelite (mudstone), locally a fine grained, clayey psammite (sandstone), with arkoses and feldspathic sandstones. Lithologically Van Hise and Leith made the same groupings as those published in Monograph 19 with the notable exception of the Quartzites and Conglomerate. These were included in the description of the Early Keweenawan. The lower contact of the Tyler was described as conformable, the upper unconformable.

A summary of the effects of metamorphism describe essentially the same lateral variance in the formation as the previous monograph. Slates, graywackes and graywacke slates found in the east became more crystalline to the west with dominantly mica slate and mica schists to the west of Penokee Gap. These alterations were attributed to the effects of contact metamorphism without addition of material from outside the Tyler Formation.

Thickness estimates were 7,110 feet and 11,000 feet for the Montreal River and the Tyler Forks sections respectively. A second cross-fault was recognized at the Potato River and horizontal offset was estimated at 280 feet.

A plot of chemical composition of a Tyler sample on a triangular diagram "clearly" showed that the Upper Middle Precambrian shale was "...the little-decomposed debris of a basic igneous rock," (Van Hise and Leith, 1911, page 612). The Early Precambrian terrain to the south of the Middle Precambrian rocks was again cited as the source area and "lagoons" and "great deltas" were favored as depositional sites for Middle Precambrian slates in general. As evidence of these environments, the authors cited numerous horizons of pyritiferous, graphitic and iron carbonate lenses within the slates.

<u>1911-1929</u>

Between 1911 and 1929 (when the Wisconsin Geological Survey published Bulletin 71), detailed work was done on the range and individual township reports were prepared by survey workers. Lake (1915a, 1915b) and the geologists of his party, C. S. Corbett, D. G. Thompson, Foster and H. D. Wakefield, payed particular attention to the Tyler, subdividing it into three members in the Penokee Gap area. These subdivisions were the upper, middle and lower members and included respectively, 3/12, 4/12 and 5/12 of the total thickness of the Tyler Formation. The upper division was extremely foliated, micaceous quartzites and graywacke slates, thickly bedded and lacking cleavage; and the lower was comprised of black, thin-bedded micaceous slates with iron carbonate and magnetite. The middle unit was slaty and carried the "unknown metamorphic mineral" which was responsible for the spotted nature of many of the slates.

<u>Hotchkiss, 1919</u>

In a series of four articles in Engineering and Mining Journal, W. O. Hotchkiss described the geology and recent mining developments on the Gogebic Range. Part of his geology was a description of the Pabst conglomerate found at the base of the Tyler Formation.

Aldrich, 1929

In 1929 all of the information gleaned from the township surveys was compiled by H. R. Aldrich and published as Bulletin 71 of the Wisconsin Geological and Natural History Survey. Several "firsts" in Gogebic Range geology can be attributed to this publication. It was the first:

1. to study glacial deposits,
2. to report cross-faults not apparent in outcrop,
3. to report jasper fragments in the Pabst Conglomerate (base of Tyler),
4. to map minor folds and at least two major ones in the Penokee-Gogebic sequence,
5. to refrain from calling the Tyler a slaty formation since it does not generally exhibit slaty cleavage,
6. to map the Keweenawan thrust fault at the base of the Mellen Gabbro.

Aldrich considered the Ironwood-Tyler contact to be a gradational one marked by only a slight time break. Although the contact does not crop out, he had access to several drill cores and to the several closely spaced outcrops at Penokee Gap and elsewhere. His studies documented the presence in the Tyler of chert and carbonate (common in the Ironwood) and a continuation of iron deposition, although on a much reduced scale, from the Ironwood up into the Tyler. Analyses for iron content in the Tyler showed 20-25% for the first 150 feet, as much as 15% for the next 150 feet and less than 1% above that, suggesting gradually diminished iron deposition. Further, he cited the presence in the Ironwood of clastic sediments. These facts were sufficient to establish to his own satisfaction that the contact was not an abrupt change from chemical to clastic sedimentation and, thus, that no time gap is evident between the Ironwood and the Tyler. An unconformity was recognized at the top of the Tyler on the basis of broad, regional relationships, but no angular discordance between the Tyler and the Keweenawan was described between Mellen and Hurley.

One of the most unique aspects of Aldrich's work was the utilization of magnetic dip needle surveys across the range. Data from these traverses enabled him to plot buried contacts, unexposed fault zones and folds, and lateral variation in formational dips between those fault zones. Thus, outlying areas to the west of Mineral Lake where no Tyler outcrops occur, were mapped as Tyler Formation on the basis of magnetic similarity to areas of Tyler outcrop. Several northwest-trending faults were located in addition to the Penokee Gap and Potato River faults previously reported. Dip of the Penokee Gap fault was determined to be westward. Successive lessening of dip from northeast to southwest of individual fault blocks bounded by the cross faults suggested to Aldrich differential foundering of the Middle Pre-cambrian rocks in post-Keweenawan time. The entire sequence was tilted by collapse of the roof of the magma chamber from which the Keweenawan lavas were extruded. The steeper dips in the northeast suggested that the collapse was of greater magnitude, or faster, or both in the northeast than in the southwest, creating a crustal warping which resulted in breaking along the faults. The resultant fault blocks tilted more in the northeast than those in the southwest. Further, he mapped a large syncline in the Middle Precambrian and lower Keweenawan rocks in the southwest corner of T44N, R4W and the southeast corner of R5W, striking N15-30^{o} East. Adjacent to and west of the syncline was an anticline striking northwest. These folds and several smaller ones in T44N, R3 and 4W, were attributed to the same differential foundering which caused the faulting.

Thrust faulting which Aldrich believed brought the Keweenawan intrusive bodies into contact with the Tyler was called upon to account for the regional beveling of the Tyler to the west. As evidence for that faulting, Aldrich cited the presence of breccias along the contact. Estimates of Tyler thickness were on the same order as those presented in Monograph 52. Several topographic highs within the Tyler Valley where numerous outcrops were found were mapped as sills and dikes, the emplacements of which were facilitated by bedding plane faults.

Metamorphic changes going westward included recrystallization of all the Middle Precambrian rocks, as evidenced by increasing grain size, and a decrease in carbonate content with a corresponding increase in magnetite, especially in the Ironwood Formation. Contact metamorphism due to the emplacement of the Keweenawan granite and gabbro was cited as the cause.

Lithologically, Aldrich made use of Lake's three units within the Tyler Formation. These subdivisions were made at the Penokee Gap section and were described as follows:

> Upper Slate Member-
> quartz and mica schists interbedded with impure quartzites or graywackes; micas present include muscovite and biotite with the possibility of sericite; beds much folded and contorted with abundant drag folds and a peculiar "boulder phase" within the schist, composed of 2-15 inch "boulders," composition same as that of the surrounding schist, schistosity bends around the "boulders" and bedding in the surrounding rock (explanation--mudballs formed prior to lithification and subsequently rotated during compaction and metamorphism);
>
> Middle Slate Member-
> gray, greenish gray, dark grayish-brown quartzitic slate, graywacke and graywacke slate; massive bedding (6 inches-4 feet); alternating with black slate containing an unknown metamorphic mineral; no schistocity; dip is uniform; metamorphism slight;
>
> Lower Slate Member-
> black, dense slate with slaty cleavage; 1/4-1 inch thick beds containing iron carbonate, magnetite and pyrite; few small drag folds and fault planes.

These members were identified largely on metamorphic and structural characteristics, rather than on original lithologies. Indeed, if one removes from the above descriptions everything but lithologic details, the Tyler appears to be quite uniformly a formation of interbedded slates and graywackes of varying bed thickness. Aldrich himself states (page 106), "...Perhaps the main body of the [Tyler] formation represents such a mixture as to warrant the name graywacke."

Gradual mineralogical change from east to west in the Tyler did play an important role in Aldrich's discussion of source area (page 106).

> "From east to west there is an apparent variation in composition which in a general way corresponds with the variation in composition of the Archean [Lower Precambrian] to the south."

From the Montreal River westward to Penokee Gap the Archean rocks are greenstone schists with intruded granites and from the gap westward they are dominantly syenitic granites. He reasoned, therefore, that sediments derived from those rocks should be dominantly graywackes in the east and arkoses in the west. This corresponded well with what he saw in the Tyler. Subsequent metamorphism altered the arkoses to mica schists.

Aldrich raises the question on page 166 of whether the Tyler sediments were derived directly from an Archean landmass or whether the weathered products of the Archean rocks were first deposited as sediment and later reworked and redeposited as the beds of the Tyler. If the latter were the case, were those first sediments lithified or were they still soft when they were eroded the second time? This paper addresses itself to those questions later (see PETROLOGY).

In many places Aldrich expresses the idea that the Tyler Formation is but a continuation of the earlier Palms sedimentation with the Ironwood representing a brief, yet prolific period of volcanic interference and onslaught of volcanically derived chert, carbonate and iron; that is, that there was no "break in the action," no cessation of deposition of

Palms clastic sediments,. but merely a comparatively fast and contemporaneous deposition of chemical sediments. This onslaught was fast enough to overshadow the clastic sedimentation still going on in the basin and resulted in a very thick, intraformational sequence of iron slate and carbonates. I consider this a very appealing hypothesis.

Regarding the tectonics and environment of deposition (pages 163-164):

> "A great area had begun to sink in early Palms time and, except for minor warpings possibly wholly within the basin itself, and having axes related to those of the basin, this sinking had persisted throughout the Upper Huronian [Middle Precambrian]. These warping movements are believed to have been related to the rising of the Great Keweenawan magma."

He goes on to say (page 166):

> "Just what is responsible for return of clastics [to the Middle Precambrian basin after Ironwood deposition] is difficult to determine, but presumably differential warpings should be considered."

Cessation of Tyler deposition was brought about by warping of the basin along an axis parallel to that basin (northeast to southwest), according to Aldrich. Beveling of the Tyler occurred in the east while shallow water conglomerates (fluvial?) and sandstones of Keweenawan age were deposited elsewhere. Aldrich records current indicators in these Keweenawan sediments and in the Palms, but none in the Tyler (see <u>Summary of Paleocurrent Indicators</u> for a compilation of indicators reported in the Tyler and adjoining formations by other authors).

Aldrich offers an explanation for the interbedded fine and coarse sediments in the Tyler. This explanation was related to impending Keweenawan volcanism. The fragmental beds (the coarser sandstones) of the Tyler, he thought, were the result of breaking up of earlier deposited beds in response to volcanic shocks and movements. The slates represent periods of quiet, stable deposition between earthquakes. In reading Bulletin 71 it is clear that Aldrich envisioned a somewhat elongated basin stretching northeast and southwest along the Early Precambrian highlands. This basin became the depocenter for this mixed bag of sediments. What is not clear, however, is how far he thought the basin extended in a northwesterly direction. Presumably, it extended at least as far as the Middle Precambrian sediments on the north shore of Lake Superior since the suggestion of correlation to those rocks was made some years earlier by Irving. Also not clear is how deep the basin must have been. At one and the same time it had to be deep enough for the normal clastic sedimentation to have been clay-sized and to remain undisturbed by wave action and yet shallow enough to receive coarse sediments loosened by volcanic shockwaves. The mechanics of this deposition were problematical.

Hotchkiss, 1933

Hotchkiss adds one more event to the summary of Tyler geology. He reported that coincident with the erosion of the Tyler and parts of the Ironwood in the east as described by Aldrich, the Presque Isle granite was intruded into the Middle Precambrian rocks.

Atwater, 1935

In 1935 G. I. Atwater finished a Ph.D. dissertation at the University of Wisconsin-Madison which was titled Correlation of the Tyler and Copps Formations of the Gogebic Iron Range. He examined the Tyler at the eastern end of the outcrop belt and the major outcrops in the west. Heretofore, even though the two formations occupy the same stratigraphic position and they are similar in outward appearance, some doubt as to the correlation of the two existed since the Tyler was lacking a basal conglomerate, one of the distinguishing features of the Copps. The Tyler does, as has been previously mentioned, have several fragmental horizons, one of which is called the Pabst Fragmental and lies at the base of the formation. The discontinuous nature of the Pabst Fragmental along strike suggested two interpretations; either there was alternating erosion and non-erosion due to geographic and topographic differences which partially removed the Pabst, or there was original non-deposition in some areas. Atwater opted for the latter interpretation since he found evidence for subaerial erosion of the underlying Ironwood Formation in abraided hematite pebbles within the Pabst. Atwater concluded that this fragmental zone was the western equivalent of the Copps Conglomerate.

Several lithologies were defined in his paper. They included graywackes, slates, massive carbonate and interbedded chert, the Pabst Fragmental, ferruginous slate and reddish slate and sandstone. The carbonate, chert and Pabst Conglomerate are to be found very near the base of the formation; the reddish slate and sandstone very near the top. Volumetrically, only the graywackes and slates were important. Atwater (1935; page 35) states:

> "The main body of the Tyler Formation consists of massive and thin-bedded graywacke, with minor proportions of interbedded dark slate."

Atwater defined three graywacke phases, a massive graywacke, a thin-bedded graywacke and a slate phase. Generally, these three were to be found interbedded but any one outcrop could be dominantly one or the other. Arbitrary selection of the dominant type in any one outcrop caused Atwater to reject them as mapable units. They proved useful only in a descriptive sense.

In thin section the massive graywacke was described as being coarse to fine grained, angular (mostly) to rounded, very crudely sorted, with quartz, feldspar and micas the dominant minerals. Accessory minerals included garnet, titanite, rutile, leucoxene, clastic chlorite (greenstone fragments?) and zircon. The matrix was chlorite, sericite and quartz. The feldspars often were altered to sericite and chlorite.

Concerning the possibility that differential erosion had introduced some bias into the exposed lithologies, Atwater stated (page 21):

> "The location of the minor hills and depressions throughout the Tyler valley does not appear to be the result so much of differential erosion on the underlying beds as it is of glacial deposition. The Tyler valley does not contain elongate

> ridges parallel to the strike of the formation, as might be expected if the slates had been eroded rapidly between massive resistant beds."

Atwater concludes that the Tyler and Copps Formations are correlative and that they belong to the Upper Huronian (Late, Middle Precambrian) separated from the Middle Huronian Ironwood Formation by an erosional unconformity.

He envisioned a gradually deepening basin (going from east to west) during Tyler-Copps deposition. He sighted as evidence of this, increasing depth of erosion to the east (indicating that the eastern area was elevated), decreasing grain size in the Pabst Conglomerate going westward (therefore, presumably into deeper water), and changing character of the Copps-Tyler sediments from Early Precambrian cobbles and boulders in the east, to greenstone fragments and finally to reworked early Huronian sediments (Bad River, Palms and Ironwood) in the west. He thought, therefore, that the basin deepened not only in a northwesterly direction away from the Early Precambrian landmass, but also in a southwesterly direction.

Leith, Lund and Leith, 1935

Leith, Lund and Leith speculated on the nature of the depocenter of the Middle Precambrian sequence. They said (page eight) that this group of rocks "...carries evidence indicating deltaic deposition."

Meek, 1935

In 1935 W. B. Meek did an M.A. thesis at Wisconsin wherein he studied the accessory minerals in the Palms Formation. He found detrital (rounded) tourmaline and rutile which suggested more than one cycle of sedimentation because of the relative mechanical and chemical stability of these minerals. He concluded, furthermore, that most of the Palms was derived from reworked Bad River dolomite and Sunday quartzite, with only the upper part of the formation having been derived from the erosion of Early Precambrian rocks, presumably due to complete removal of the lower Middle Precambrian sediments. Tectonically, this suggested that the basin which served as the depocenter for the Tyler Formation was already present during Palms time. If the Upper Palms sediments were derived from Early Precambrian rocks as he suggests, then the Tyler sediments also had only Early Precambrian rocks for a source, since the Lower Middle Precambrian sediments must have been completely removed during Palms deposition.

Tyler, Marsden, Grout, and Thiel, 1940

In a definitive work on the heavy mineral suites of the Precambrian rocks of the Lake Superior region, Tyler, et al. devised a scheme of age identification based upon zircons. Briefly, pre-Huronian igneous rocks contained purple (hyacinth) zircons; Huronian, pre-Keweenawan igneous rocks contained weakly birefringent (malacon) zircons; and Keweenawan ("normal") zircons were euhedral, and colorless to yellow.

Dates of unroofing of the batholiths containing these species was reflected in the sediments derived therefrom. Thus a sediment containing yellow zircons, for example, could be no older than Keweenawan. The zircons in the Tyler (along with the other sediments of the Penokee series) dated the Tyler at post-Archean.

Williams, Turner and Gilbert, 1954

The authors of this petrography text cited a sample of the Tyler Formation as one example of a graywacke even though it lacked "significant" rock fragments. Its texture and chlorite-rich matrix, they felt, were sufficient criteria to classify it as such.

Hendrix, 1960

A dissertation on the structural aspects of the east Gogebic Range by T. E. Hendrix in 1960 described what is known as the Presque Isle Volcanics (see Fig. 2). They consist of flows, flow breccias, agglomerates and conglomerates deposited while Middle to Late Middle Precambrian sediments were being laid down in the west. These volcanics interfinger with the Ironwood Iron Formation and immediately underlie the Copps Formation. It is possible that these volcanics may have contributed sediments to the Tyler Formation.

Prinz, 1967

Mapping by W. C. Prinz in Michigan showed that the Lower Tyler and Upper Ironwood are interbedded with mafic volcanics in the eastern region. To date, neither volcanics nor sediments derived therefrom have been reported in the Tyler on the Wisconsin side.

Cannon and Gair, 1970

The name Marquette Range Supergroup was proposed in 1970 to include the Paint River, Baraga, Menominee, and Chocolay Groups in Michigan (see Table 1) in an effort to supplant terms like Animikie and Huronian. Use of such terms, according to the authors, suggested unproven correlation with those rocks in Minnesota and Canada. The Tyler Formation presently is considered part of the Baraga Group (U.S.G.S.).

Felmlee, 1970

J. K. Felmlee studied the structural aspects of the Middle Precambrian-Late Precambrian contact near Mellen, Wisconsin. Although the emphasis of her work was on the nature of the contact, she studied the sedimentology of the Tyler to a certain extent. She discovered graded bedding, cross-laminae, ripple-drift and load casts in the formation. She states that the currents which built the cross-laminae and ripples were probably not related to the currents which deposited the sediments. The paleocurrent measurements which she made indicate flow toward the northwest (see Summary of Paleocurrent Indicators). It should be emphasized, however, that she believed these currents were normal bottom currents which reworked the sediments and not the same currents which initially deposited them. Implicit in her remarks is the belief that turbidity flows were the mechanism of transport and deposition, but she describes no sole marks on the beds. She called the Tyler-Lower Keweenawan contact a disconformity.

Previous authors called upon the Keweenawan gabbroic and granitic intrusions near Mellen to account for increasing metamorphic grade in the Middle Precambrian sediments to the west. Felmlee, on the other hand, thought the distances were too great between the intrusive rocks and the sedimentary rocks affected, and attributed the changes to regional metamorphism. That is not to say there is no contact metamorphism around the gabbro; she mapped a zone up to 1/2 mile wide of hornblende-hornfels and pyroxene-hornfels facies. Much of the Tyler, especially near intrusive bodies, was mapped as schist.

Felmlee mapped a large scale fold in the Tyler in R2W based upon differences in dip between the northeast-southwest striking beds and the more easterly striking beds. She also mapped another cross-fault, this one trending northeast, at Loon Lake, and three sets of diabase dikes and sills within the Tyler. The faults and dikes, she stated, probably developed during the formation of the Lake Superior syncline in Keweenawan time. She concluded that the Tyler was probably not folded during what has come to be known as the Penokean Orogeny (originally defined at 1.7 b.y. by Goldich, et al., 1961) and that it is improbable that the Keweenawan thrust fault passes through the area at the Middle Precambrian-Keweenawan contact, or at the base of the Mellen Gabbro.

Trent, 1971

Investigations by the U. S. G. S. in the east Gogebic Range area are summarized in this Professional Paper (750-A) and suggest a shallow water, marine environment possibly associated with an island arc as the depositional environment for the Middle Precambrian rocks. Evidence cited includes pillowed lavas, ripple-marked bedding planes, algal structures and scour and fill. An island arc environment is suggested by the similarity of the rock sequence with that of modern island arc systems.

Komatar, 1972

Komatar did an M. S. thesis at Madison, Wisconsin entitled Geology of the Animikian Metasedimentary Rocks, Mellen Granite and Mineral Lake Gabbro west of Mellen, Wisconsin. His study included all of the Middle Precambrian sequence, the gabbro, granite and lower volcanics.

According to Komatar, the Bad River Formation was deposited unconformably on the Early Precambrian granitic terrain which undoubtedly formed a stable marine shelf. Carbonate and quartz sands with algal structures indicated shallow water. The depocenter was apparently being slowly filled with sediments as evidenced by increased clastic content upward in the formation. Komatar described the Bad River-Palms contact as a disconformity.

The Palms Formation is comprised of three units, according to

Komatar; a basal conglomerate, an argillaceous middle unit and a clean quartzite at the top. He found small (1/8-3/8" diameter) mafic volcanic pebbles in the Palms. Primary sedimentary features included cross-beds and sole marks indicating currents from the southeast (see <u>Summary of Paleocurrent Indicators</u>). He recognized a general directional change in the currents in the Palms with the current indicators in the lower beds being southerly and the upper indicators more easterly. The environment of deposition was envisioned as a shallowing basin which achieved tectonic stability in upper Palms time resulting in a pause in deposition. Those sediments already deposited were reworked to a mature quartz sand. The upper contact of the Palms with the Ironwood was described as conformable. The transition from Palms clastics to Ironwood chemical precipitation took place within one inch stratigraphically, a drastic change which Komatar found difficult to explain.

The Ironwood, according to Komatar, represented a period of chemical precipitation in a restricted basin. Ripple marks of very small wavelength (1/3-1/5") suggested currents flowing from S50^{o}E (see <u>Summary of Paleocurrent Indicators</u>). He described the Ironwood-Tyler contact as conformable and gradational.

The Tyler was described as an argillaceous sandstone or subgraywacke according to the definition of Moorhouse, (1959). He noted the following flysch-like characteristics:

1. rapid alteration of fine shales and coarse sandstone,
2. sharp basal bedding contacts with gradational upper contacts,
3. graded beds,
4. sole marks.

Besides those features, he found cross-bedding and rip-up clasts. Average bedding thickness was determined to be three inches with great variance.

Petrologically, Komatar gave an average composition for the Tyler argillaceous sandstones as 53% quartz, 24% hornblende and biotite, 9% muscovite and chlorite, 11% plagioclase, and 3% opaques. His description was of metamorphosed rocks at the western end of the outcrop belt. The original Tyler was undoubtedly much richer in feldspars and deficient in hornblende, micas and chlorite. Many of the quartz grains were "embayed" but whether this suggested to Komatar original volcanic quartz or embayment due to metamorphism is unclear. He also mapped several drag folds within the Tyler.

The Tyler was interpreted as having been deposited in a deepening marine environment. The rapid change from chemical to clastic deposition and the flysch-like character of the beds suggested a period of rapid uplift and subsequent erosion of the source area to the southeast. Komatar recorded paleocurrent indicators in the Tyler which suggested movement toward the NNW (see <u>Summary of Paleocurrent Indicators</u>).

The Penokean Orogeny, Komatar thought, was characterized in the Tyler region by warping and minor folding resulting in regional metamorphism to the greenschist facies. Subsequent erosion and deposition of the overlying Keweenawan Bessemer Formation resulted in an angular unconformity at the top of the Tyler. Following the Penokean Orogeny and deposition of the Bessemer, simultaneous volcanism and tilting toward the north began.

Mattis, 1972

In an M.S. thesis at the University of Minnesota-Duluth Mattis described the Tyler-Bessemer contact as a disconformity. Currents in the Keweenawan Bessemer basin flowed SW according to Mattis (see Summary of Paleocurrent Indicators).

Schmidt, 1972

Schmidt mapped the Precambrian rocks of the Ironwood-Ramsay area of Michigan and made this statement (page 8) with regard to Tyler outcrop distribution:

> "Fine quartzite is the most common rock in the natural outcrops [above the lower ferruginous zone], but that is because the argillite erodes more easily."

He estimated maximum thickness to be 9500', and described syngenetic sulfides in the Tyler, an indicator of reducing conditions during deposition.

Schmidt and Hubbard, 1972

The Eighteenth Annual Institute on Lake Superior Geology sponsored a field trip to the Gogebic region in 1972 for the purpose of examining the structure of the region. Several stops were made to examine the Tyler Formation and a portion of the guide book text reads as follows (page A19):

> "Because the progressive upward increase in dip of the Precambrian X [Middle Precambrian] and lower Keweenawan rocks is a regional relationship, it is inferred that the differences in dip are related to deposition of the rocks and that the central parts of the depositional basins of both Precambrian X and Lower Keweenawan rocks were to the south. The center of the depositional basin of the Middle and Upper Keweenawan rocks was to the north."

Cooper, 1973

Cooper studied the Middle Precambrian and Keweenawan rocks in the Mellen-Hurley area. He described the Ironwood-Tyler contact as an erosional unconformity and the Tyler-Bessemer contact as an angular unconformity. He found cross-laminae, graded bedding, current ripple marks (some sand-starved ripples) and imbricated mud chips in the Tyler. On the basis of mud chip imbrication and flame structures (personal communication) he suggested north to south movement of the currents which deposited the Tyler sediments. The rocks were described as immature sandstones and slates. Quartz amounted to 50-60% of the total volume, feldspars were estimated at 5-10%, biotite 1-5% and variable amounts of chlorite, muscovite and opaques.

Cooper envisioned a topographic high to the north during Late Middle Precambrian and possibly to the east during Early Keweenawan.

The interstratification of the graded graywackes and slates in the Tyler Formation was the result of periodic movement of coarse sediments from shallow water surrounding the landmass into deep water where the normal sedimentation was pelagic muds. The mechanism of transport and deposition, he thought, was turbidity currents.

STRUCTURE AND THICKNESS

The stated objectives of this paper do not include a close examination of the structure of the Tyler. Much previous work by several different people has been done with the express purpose of establishing a structural history of the Gogebic Range area. The results of a few of those studies are presented in Table 2. During the course of field work, however, several observations were made and are recorded here in the hope that they will benefit those particularly interested in the local or regional structure.

As stated in the GEOLOGIC SETTING section of this paper and elsewhere, the Tyler Formation, like all of the Middle Precambrian sequence, presently dips steeply toward the northwest and trends northeast-southwest. It forms a structural monocline, a part of the southern limb of the Lake Superior Syncline. Bedding attitudes were consistent with but few exceptions (Fig. 5). Those anomalous readings were always explicable by local slumping, intrusive bodies or faults.

Felmlee (1970) mapped a large, westward plunging flexure in the Tyler Formation in R2W on the Wisconsin side. Her observation that the east-west striking beds tend to be more closely vertical than the northeast-southwest striking beds suggested the presence of a fold. My own data for that area is generally consistent with her observation but paucity of outcrop, I feel, makes the proposal of a fold a tenuous one. I do not have enough data in my own notes to eliminate the

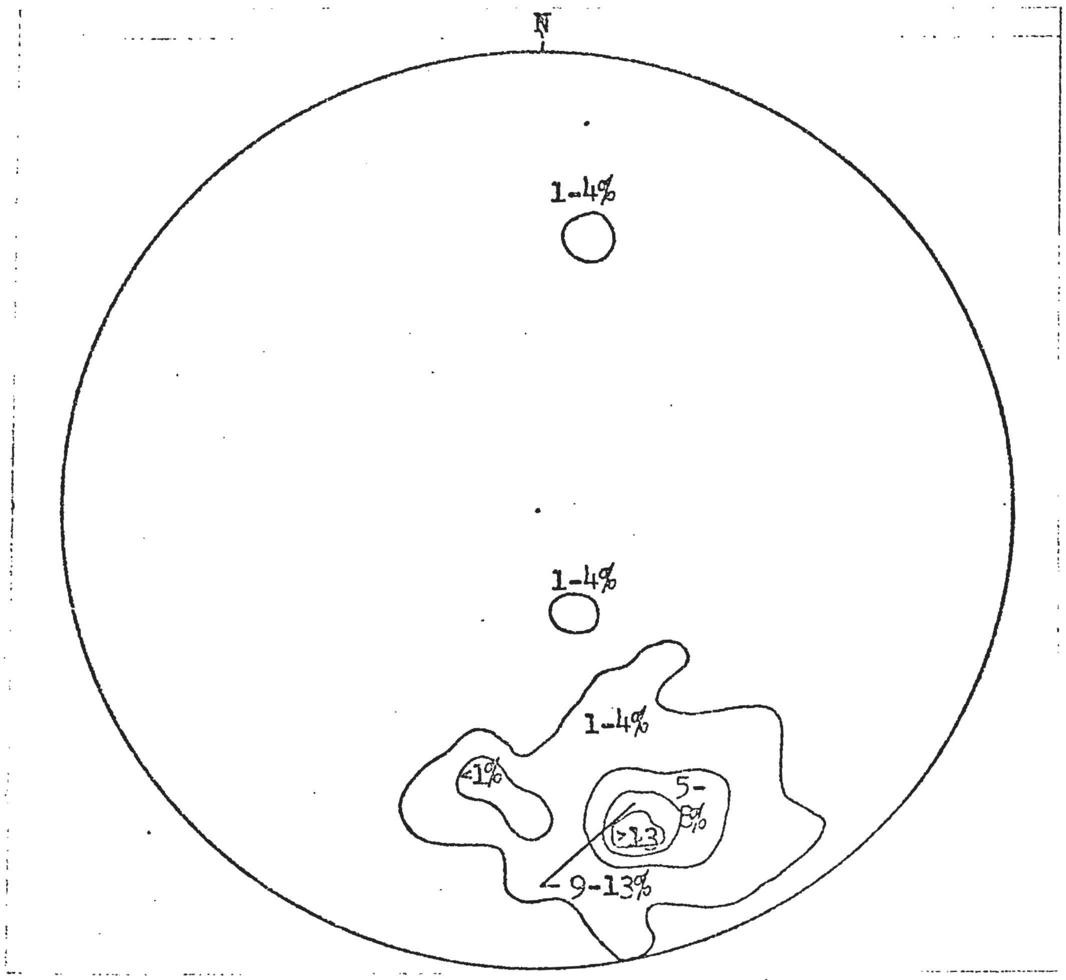


Figure 5--STEREONET PLOT OF POLES TO BEDDING PLANES MEASURED IN THE TYLER FORMATION. Total of 108 measurements. Contour interval 4%.

possibility that the fold evidence may, in fact, be due to local structural abnormalities caused by faulting and/or small intrusive bodies.

Komatar (1972) mapped four folds in the Tyler Formation in T44N, R3W and T44N, R4W. Their axes' attitudes are as follows: N67°W, plunging 10°W; N87°W, plunging 15°W; N89°W, plunging 15°W; N88°W, plunging 25°W.

A few outcrop-size folds were mapped during the present study: two small synclinal folds on the west bank of the Bad River, NE 1/4, SW 1/4, S11, T44N, R3W the axes of which strike 110°-290° and plunge 33° and 25° to the northwest; an anticline and accompanying syncline at NW 1/4, NW 1/4, S9, T45N, R1E with an axis orientation of 45°-225° plunging 50° to the northwest; a flexure (antiform) in almost vertical strata which has left the beds in the upper part of the outcrop overturned, in the southeast corner of NE 1/4, NE 1/4, S29, T45N, R1W; and a similar "S-shaped" fold in an outcrop at NE 1/4, NW 1/4, S34, T45N, R2W. It is possible that the two folds on the Bad River are the same as those mapped by Komatar.

Numerous examples of slickensides, joint patterns, small faults with displacements measureable in centimeters, quartz veins and igneous dikes were recorded, but no detailed study was undertaken.

Large scale cross-faults trending northwest and northeast (Aldrich, 1929; and others) which cut the entire Middle Precambrian sequence and the lowermost Keweenawan rocks are, for the most part, not recognizable within the Tyler Formation alone. Due to the internally consistent lithology such faults must be recognized by offset of the Tyler-Bessemer or Ironwood-Tyler contacts and traced into the area, or located by means of magnetic survey.

At the eastern end of the study region (especially near Hurley, Wisconsin) the less competent beds of the Tyler exhibit a well-developed slaty cleavage (Fig. 6). Explanations of the development of the cleavage

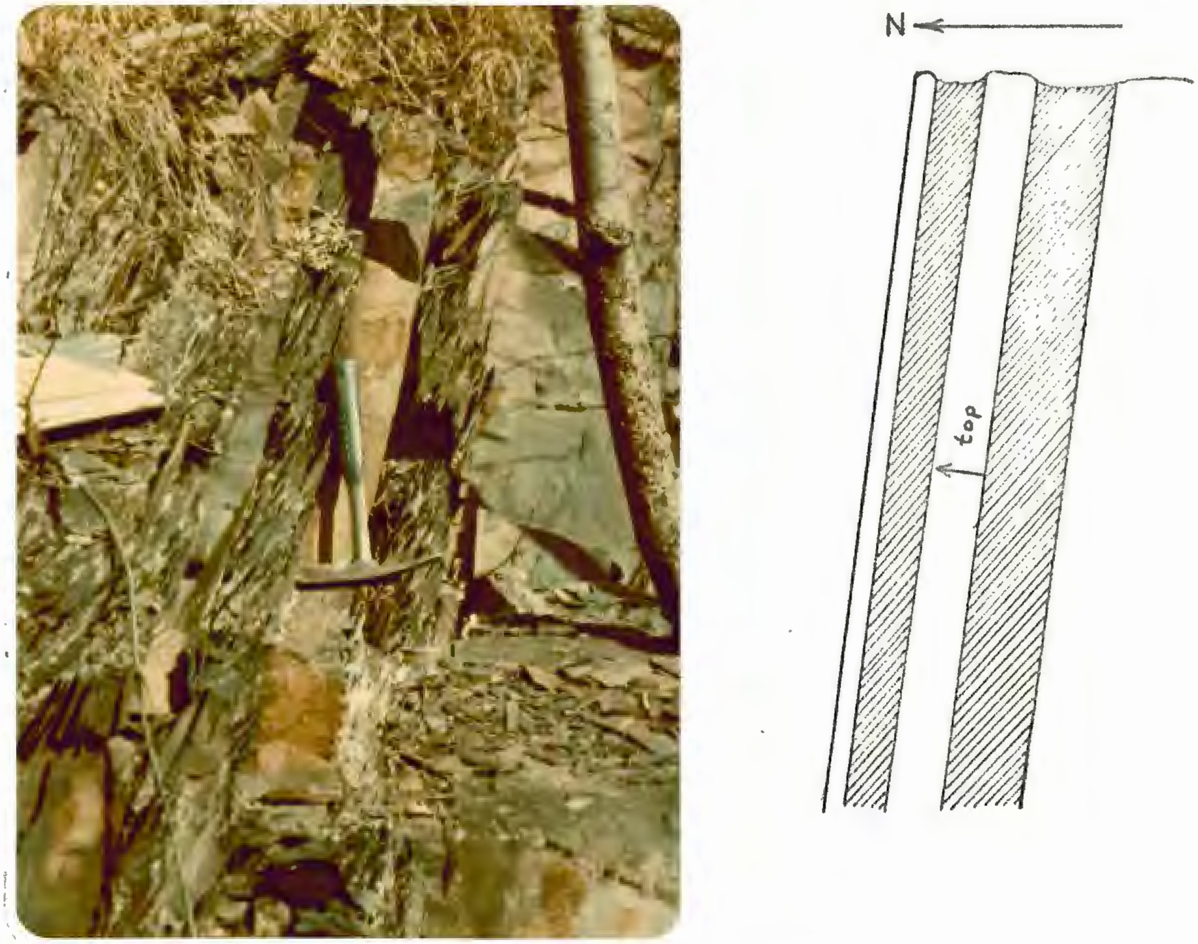


Figure 6--PHOTO AND DIAGRAMMATIC CROSS-SECTION OF INCLINED STRATA NEAR HURLEY, WISCONSIN SHOWING THE ATTITUDE OF SLATY CLEAVAGE DEVELOPED IN SLATE BETWEEN MORE COMPETENT GRAYWACKE BEDS. Note that the cleavage dips less steeply than the beds and that the beds are not overturned.

must be consistent with these facts: both the beds and the cleavage dip to the northwest but the cleavage dips less steeply than the beds; topping direction in the beds is to the northwest as shown by graded bedding and cross-stratification.

The cleavage cannot be axial plane cleavage developed in response to the same forces which tilted the Tyler to its present attitude.

If it were, it would be inclined more steeply than the strata which contains it (Fig. 7a), or, the strata would be overturned (Fig. 7b). It cannot be fracture cleavage for there also, the cleavage dips more steeply than the strata (Fig. 8).

There are three possible explanations. The cleavage may be flow cleavage developed subsequent to folding (tilting) of the Tyler and as a result of a secondary force caused by a couple, S-S', as shown in Fig. 9. Alternatively, the cleavage may be axial plane cleavage developed on a refolded fold (Fig. 10). A third hypothesis was stated by Cooper, (1973, p. 22):

> "The only apparent solution to this problem is that the beds at the time of deposition dipped to the south and that the cleavage formed shortly after deposition. Subsequent to the formation of the cleavage the beds were rotated through 90 degrees to their present attitude."

This hypothesis is a variation of that shown in Fig. 9.

The relative merits of each of these three hypotheses cannot be evaluated here. Suffice it to say that there is ample evidence for two separate periods of tectonic activity (see Table 2) which the first two hypotheses require and also for southerly dip as required by Cooper's hypothesis (see Hendrix, 1960).

Table 3 is a tabulation of attitudes of slaty cleavage and bedding at a few outcrops. Those figures point out the fact that for the limited number of observations made in conjunction with this study, the dip of the slaty cleavage remains relatively constant from outcrop to outcrop

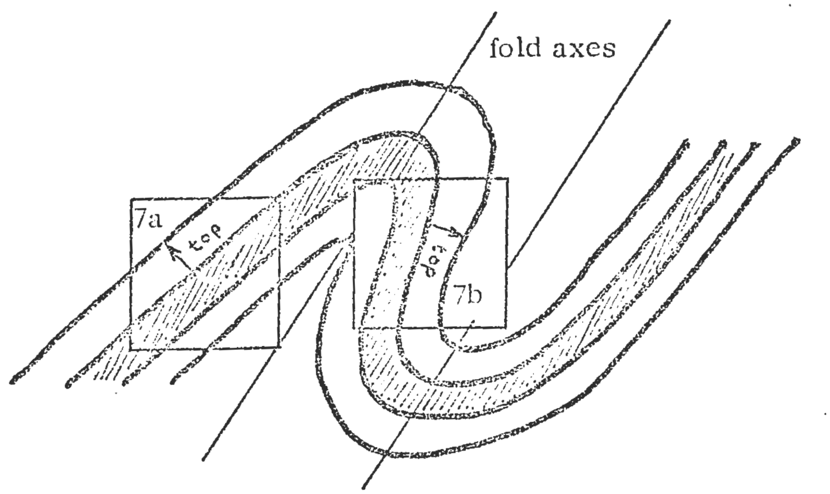


Figure 7--SKETCH OF FOLDED STRATA WITH WELL DEVELOPED AXIAL PLANE CLEAVAGE. Note that in 7a the cleavage dips more steeply than the strata and in 7b the beds are overturned. Contrast Figures 7a and 7b with Figure 6. (After Billings, 1972)

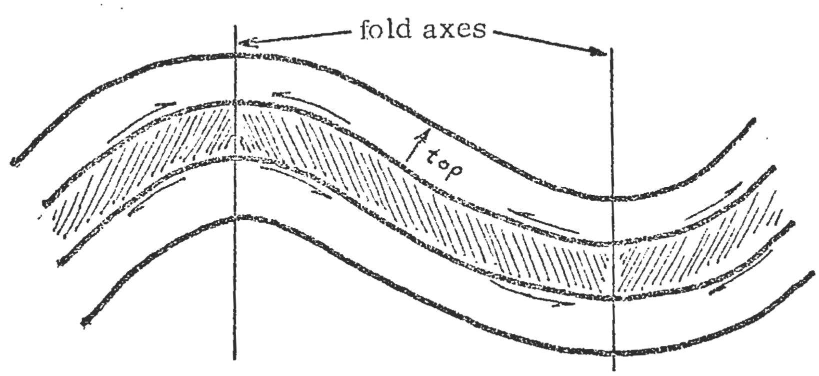


Figure 8--SKETCH OF FRACTURE CLEAVAGE DEVELOPED AS A RESULT OF SLIPPAGE ALONG BEDDING PLANES DURING FOLDING. Note that the cleavage dips more steeply than the strata. (After Billings, 1972)

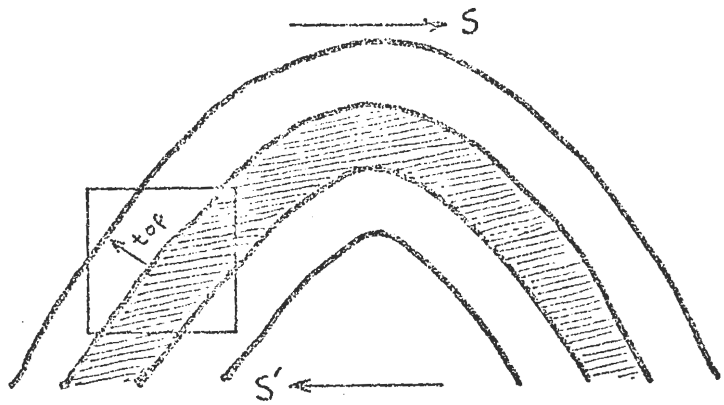


Figure 9--SKETCH OF FOLDED STRATA WITHIN WHICH SLATY CLEAVAGE HAS DEVELOPED IN RESPONSE TO THE FORCE COUPLE S-S' AFTER INITIAL FOLDING. Compare the boxed area to Figure 6. (After Billings, 1972)

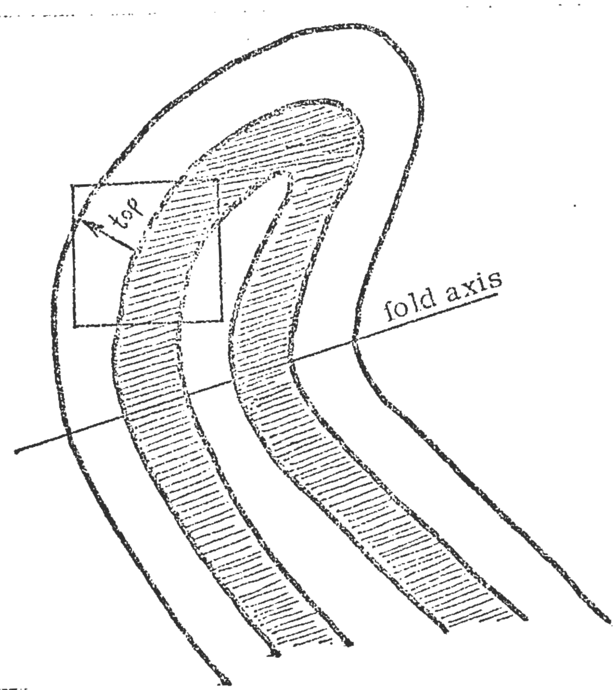


Figure 10--SKETCH OF A REFOLDED FOLD WITH SLATY CLEAVAGE DEVELOPED IN THE INCOMPETENT STRATA. The cleavage is parallel to the axial plane of the latest folding. Compare boxed area to Figure 6. (After Billings, 1972)

TABLE 3--ATTITUDES OF SLATY CLEAVAGE AND BEDDING AT A FEW OUTCROPS

Outcrop number*	Slaty cleavage (degrees)		Bedding (degrees)	
	strike	dip	strike	dip
TS 45-32....	71	54N	71	68N
TS 45-245...	77	55N	77	71N
TS 45-41....	73	66N	74	76N
TS 41-10....	72	53N	72	62N
TS 43-14....	77	54N	80	83N

*see Plate I for outcrop locations

TABLE 4--COMPARISON OF THICKNESS ESTIMATES, TYLER FORMATION

Source	Penokee Gap (feet)	Tyler Forks River (max.) (feet)	Montreal River (feet)
Irving, 1880.......	11,480		
Irving & Van Hise,. 1892		12,360	7150
Van Hise & Leith, . 1911		11,000	7110
Aldrich, 1929		10,000	
Schmidt & Hubbard, 1972		9500	7000
Cooper, 1973		10,000	
This paper........		12,000	7900

while dip of the strata vary considerably. If more data were to confirm these few measurements it would suggest that the cleavage was imprinted over the strata after they achieved their present attitudes.

The Tyler Formation is by far the thickest unit of the Middle Precambrian sequence. Estimations of its thickness are based on the assumption that the Tyler is a simple homocline and that there is no repetition of beds due to internal folding or faulting. No evidence of repetition of bedding was found in the course of this study. By simple trigonometric calculations using the outcrop belt surface width perpendicular to strike and the average dip of the beds, the Tyler is estimated to be 12,000 feet thick at its widest point (Tyler Forks area). Near Hurley, where the stratigraphic section was measured, the thickness is estimated at 7900 feet. These figures are compared with those of other authors in Table 4. There is, of course, no way to measure the thickness of Tyler Formation removed by erosion prior to deposition of the Bessemer Formation.

The Tyler thins and finally pinches out entirely to the west due to thrusting of the Keweenawan rocks southward and into contact with the older rocks (Aldrich, 1929; and Dutton and Bradley, 1970). Felmlee (1970), Tabet (1974) and others suggest an alternative explanation. They found no evidence for a fault in the Mellen area and suggested that the contact is probably of an intrusive nature, emplacement of the Mellen Complex having been facilitated by a plane of weakness which was the post-Tyler erosional unconformity. To the east, the Tyler gradually thins and disappears as a result of erosion before Bessemer deposition (Schmidt, 1972; Schmidt and Hubbard, 1972). In that area the Tyler has been faulted prior to erosion.

STRATIGRAPHY AND LITHOLOGY

As a direct result of economic interest in the Ironwood Iron Formation and due to relative structural simplicity, the stratigraphic succession in the Gogebic Range Area has been well known for almost a century. Good summaries of the Middle Precambrian stratigraphy of the present study area can be found in Irving and Van Hise (1892), Van Hise and Leith (1911), Aldrich (1929) or, more recently, Komatar (1972), Schmidt and Hubbard (1972). Table 1 of this paper is a general stratigraphy and correlation summary. The local stratigraphy is discussed in detail under the heading GEOLOGIC SETTING.

Since the earliest studies by R. D. Irving (1877a) the only changes wrought in the Middle Precambrian succession have been those made in nomenclature, and they have been few. Specifically, the Tyler Formation has been called successively the Upper Slate Member of the Penokee Series (Irving and Van Hise, 1892) and the Tyler Slate (Van Hise, 1901). The Lexicon of Geologic Names (Keroher, et al., 1960) presently uses the name Tyler Slate. I feel that that usage is not consistent with the known geology of the Tyler in that it incorrectly suggests a dominant lithology. Throughout this paper the youngest Middle Precambrian formation of the Gogebic Range area will be referred to as the Tyler Formation.

Near Hurley, Wisconsin there are several outcrops which show sufficient detail and which lie near enough to one another to facilitate the establishment of the most complete stratigraphic section possible in the Tyler Valley. The locations of those seven outcrops are plotted on Plate I. By projecting the tops and bottoms of those separate columns along strike, a single representative section of the Tyler Formation is established and has been reproduced in Plate II. The measured thickness of that column is 347 m. and represents approximately 10% of the total estimated thickness of the Tyler at that location. Table 5 summarizes the measurements made at and between each outcrop in a complete section.

TABLE 5--ESTIMATES OF THICKNESSES OF COVERED INTERVALS AND MEASURED THICKNESSES AT AND BETWEEN SEVEN OUTCROPS NEAR HURLEY, WISCONSIN

Interval	Estimated thickness of covered interval (m.)	Measured thickness of outcrop (m.)
top of TS45 to top of Tyler Formation	100	
TS45		210
top of TS44 to base of TS45	1760	
TS44		22
TS43B		52
TS43		4
TS41		14
TS40		13
top of TS46 to base of TS 40	610	
TS46		53
base of Tyler Formation to base of outcrop TS46	630	
total covered	3100 m	total exposed 368 m

total thickness 3468 m

A total of 1703 beds were measured pursuant to the erection of the representative section of the Tyler (Table 5). All beds are clastic sedimentary rocks and three distinct lithologies are present--sandstone, siltstone and shale--in varying degrees of metamorphism. This study focused upon those rocks exhibiting little or no metamorphism.

For convenience, the three lithologic types have been modified slightly in the following discussion. Of the total 1703 beds measured, 701 (41%) are argillites or slates, 412 (24%) are sandstones (graywackes) or siltstones which exhibit sedimentary structures indicative of turbidity current deposition (see PRIMARY SEDIMENTARY STRUCTURES) and 590 (35%) are graywackes and siltstones which lack significant primary sedimentary structures. These three lithogenetic subdivisions overlap each other in size of clastic grains and the divisions have instead been made on the basis of the presence or absence of recognizable primary sedimentary structures. Certain sedimentary structures, which will be described later, are thought to be indicative of specific mechanisms of sediment transport and deposition. Thus the subdivisions which have been made are not altogether descriptive in nature, but are to some extent interpretive.

The same lithogenetic breakdown has been made for each of the seven outcrops which comprise the measured section and the figures are presented in columns 2, 4, and 6 of Table 6.

With only three exceptions, argillaceous rocks make up the greatest percentage of total number of beds in each outcrop (column 2).

TABLE 6--PERCENTAGES[a] OF EACH OF THE THREE LITHOGENETIC SUBDIVISIONS FOR EACH OUTCROP IN THE MEASURED SECTION

OUTCROP	ARGILLITE AND SLATE		STRUCTURELESS SANDSTONES AND SILTSTONES		TURBIDITES (graded, laminated, etc.)			
	1 % of total thickness	2 % of total # of beds	3 % of total thickness	4 % of total # of beds	5 % of total thickness	6 % of total # of beds	7 columns 3 + 5	8 columns 4 + 6
top								
TS45-upper roadcut ...	12	44	68	37	20	19	88	56
TS45-lower roadcut ...	12	36	25	26	63	38	88	64
TS44...	21	42	74	44	5	4	79	48
TS43B...	24	41	47	40	29	19	76	59
TS43...	40	43	25	36	35	21	61	57
TS41...	42	44	46	48	12	8	58	56
TS40...	44	52	26	36	30	12	56	48
TS46...	44	43	34	36	22	21	56	57
base								
percentages for entire measured section	22	41	38	35	40	24		

[a]Columns 1, 3, 5 and 7 give percentages based upon volumetric abundances (thickness). Percentages recorded in columns 2, 4, 6 and 8 are based upon numerical abundances.

And in those three cases the difference in percentage points between the most numerous lithology and the argillaceous beds is small (4% or less). In all cases but one (TS45-lower roadcut), the structureless sandstone and siltstone beds are more numerous than the turbidites.

There seem to be no significant trends in the numerical importance of the lithogenetic subdivisions going up section. Therefore, subdivision of the Tyler Formation into separate members on the basis of lithology is impossible. Any single outcrop exhibits any and all of the three lithogenetic units.

Also included in Table 6 is an evaluation of the volumetric importance of the three subdivisions for each outcrop expressed as a percentage of total thickness (columns 1, 3 and 5). Those figures suggest that although argillite and slate beds are usually the most numerous, structureless sandstones and siltstones volumetrically are more important in five outcrops (TS41, TS43B, TS44, TS45-lower roadcut and TS45-upper roadcut). In three of those outcrops (TS43B, TS45-lower roadcut and TS45-upper roadcut) both structureless sandstones and siltstones and turbidites make up more of the total volume than argillaceous beds. In outcrop TS45-lower roadcut, the turbidites occupy first place. In fact, the volumetric percentages for each type for the entire measured section (bottom line, columns 1, 3 and 5) suggest that turbidites and sandstones-siltstones are of equal importance, with argillaceous beds coming in third.

There is a significant trend which should be noted in the thickness figures for argillaceous beds. Those beds seem to make up progressively less of the total volume of each outcrop up section. The bar graphs in Figure 11 show that the argillite and slate beds get progressively thinner up section. That is, the percentage of argillaceous beds which are ten centimeters or less in thickness increases steadily from TS40 to TS45-upper with but two anomalous outcrops. What is the significance of this trend?

Assuming that the rain of pelagic sediment in the basin of deposition proceeded at a constant rate, the thinning upward trend in the argillaceous rocks suggests that the events which fostered deposition of the interbedded sandstones and siltstones became progressively more frequent as time passed. If these events were more frequent, then any given outcrop which occupies a higher position stratigraphically than a second outcrop should have a greater numerical (<u>not</u> thickness) percentage of coarser clastics (sandstones, siltstones and turbidites) than does the second. Reference to column 8 of Table 6 shows that this is <u>not</u> the case. It follows, therefore, that the initial assumption of a constant rate of pelagic sedimentation is incorrect.

The alternative explanation is that the rate of fallout of clay-sized sediment in the deep waters of the basin decreased with time. Such a decreasing rate of fine sedimentation could be explained in several ways.

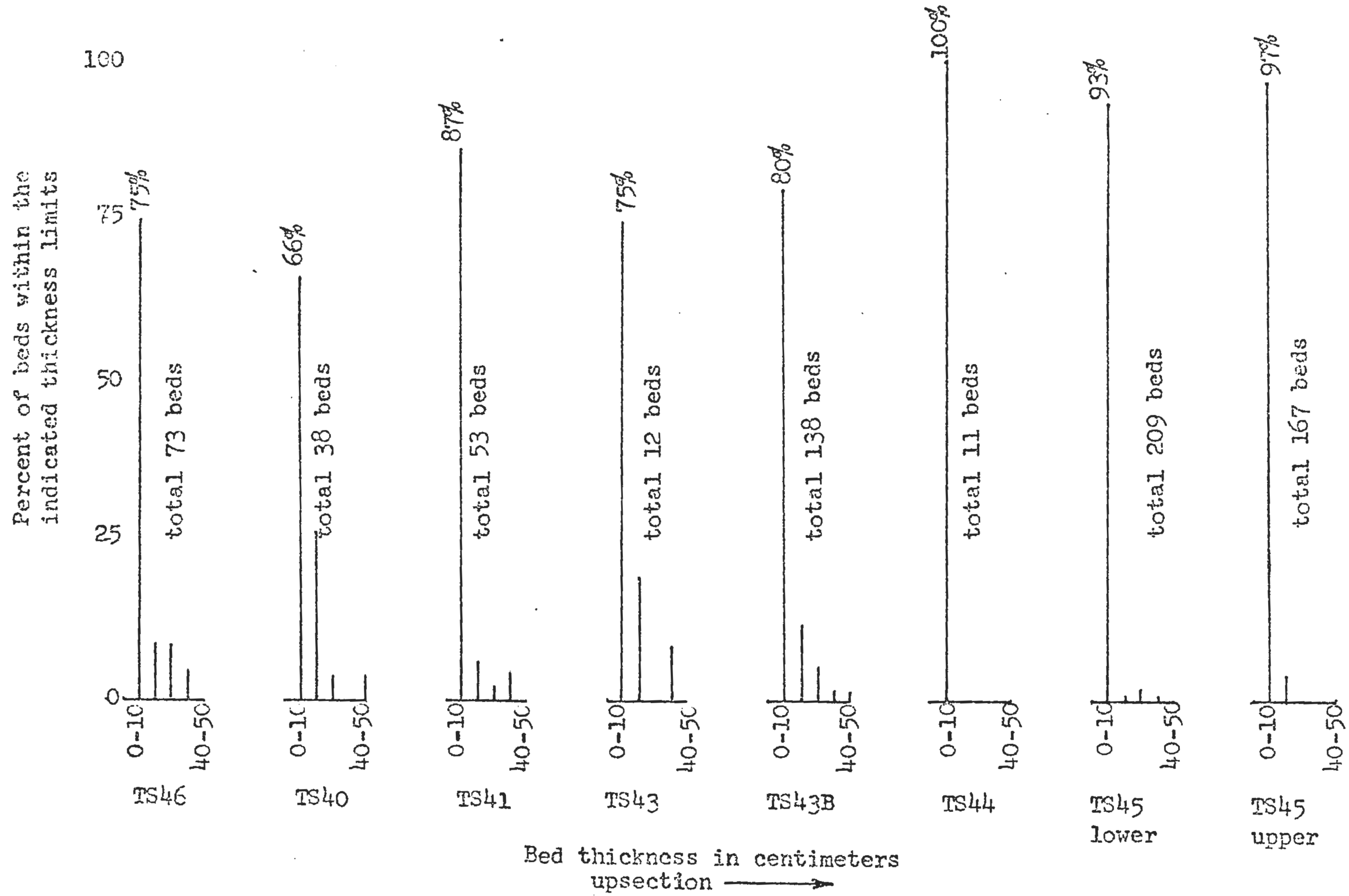


Figure 11 --VARIATION IN THICKNESS OF ARGILLACEOUS BEDS WITH STRATIGRAPHIC POSITION

The rate of sedimentation of pelagics in the depocenter depends directly upon the rate of supply of those sediments to the waters of the basin by streams draining the source area. (Relatively short-lived events in the basin itself which would effect the rate of settling of fine grained sediments, like agitation due to storms, are ignored.) The sediment loads and discharges of streams are controlled by four variables; relief, climate, vegetation and bedrock geology.

Since the Tyler Formation is Precambrian, the vegetation variable can be eliminated. Paleocurrent reconstruction (see Paleocurrent Indicators) suggests no variation in current orientation up section. That is, the currents which deposited the youngest sediments of the Tyler were moving in the same direction as those which deposited the oldest sediments. They were, therefore, draining the same source area, i.e. the same bedrock geology. There is no suggestion that the geology of the source area changed significantly (for example, by massive intrusions) during deposition of the Tyler Formation. For these reasons, the bedrock geology variable can also be eliminated. Lack of vegetation during Middle Precambrian time and lack of much geophysical data needed to establish pole locations before, during and after Tyler deposition make evaluation of climatic influence on sediment discharge rates impossible at this time. Therefore, of the four variables, only the effect of relief can be considered.

In general, the greater the relief the greater the rate of denundation by erosion and the greater the sediment discharge. And the single most important control on relief is tectonic activity. Therefore, by reverse reasoning, a decreasing rate of sedimentation up section suggests a decreasing rate of denundation brought about by a gradual lessening of local relief. Tectonically, this reasoning suggests increasing relative stability between the depocenter and the source area with time. The mechanism by which this stability was achieved (cessation of basin subsidence, peneplanation of source area or filling of basin by sediments) is unknown.

There are, of course, alternative explanations for the thinning-upward trend of the mud units. If the shoreline was retreating (i.e. transgressive sequence) during Tyler time as it seems to have been throughout Middle Precambrian time (see Table 1), then beds stratigraphically higher in the Tyler would have been deposited in progressively deeper water. And in general, beds deposited in deep water are not thick. The thinning-upward trend may also be explicable by the lateral migration of submarine channels or some other phenomena not yet fully understood.

Table 6 points out another trend which should be noted. Whereas total number of sandstone and siltstone units (including turbidites) shows no consistent variation up section (column 8) there is a definite increase in volumetric importance of these lithologies up section (column 7). It follows that the coarse units must be getting thicker with time. This phenomenon can also be seen in Plate II. But, reference to Table 7 will show that the average thicknesses of beds of each lithogenetic type

TABLE 7--AVERAGE THICKNESSES OF BEDS OF EACH OF THE THREE LITHO-GENETIC TYPES FOR EACH OUTCROP OF THE MEASURED SECTION

outcrop	argillites and slates (cm)	structureless sandstones and siltstones (cm)	turbidites (cm)
top of section			
TS45-upper roadcut	2.7	15	14.1
TS45-lower roadcut	6.6	17.8	32.4
TS44	4	11.1	12
TS43B	7.8	14.6	22.3
TS43	8.3	6.3	14.4
TS41	7.7	8.3	11.9
TS40	8.5	6.6	26.1
TS46	23.3	19.8	23
base of section			

show no significant trend up section. Therefore, the trend seen in column 7 of Table 6 must reflect the influence of a few very thick sandstone beds. In a statistical sense, the range of bedding thicknesses of the coarser grained units has increased while the mean has remained relatively constant.

An abbreviated explanation of this invasion of thicker beds up section will be offered here and explained more fully later (see <u>Discussion of Bedding Features</u>). It is thought that the few very thick sandstone beds at the top of the section are lithified channel-fill deposits which mark the locations of laterally migrating submarine fan distributary channels.

At this point it should be noted that in this study, the figures presented in the previous tables and the discussion of primary sedimentary features which follow, are based upon that part of the Tyler which is volumetrically most important. Unique lithologies near the base of the formation, like highly ferruginous slates, cherts and carbonates, were not studied in detail for these reasons: there is a general lack of outcrop near the base of the formation and where outcrops are found (e.g. Penokee Gap) the rocks are metamorphosed; the special rocks have been studied before (see Aldrich, 1929; and others) whereas that great monotonous pile of slates and graywackes which comprises the bulk of the formation has not.

PRIMARY SEDIMENTARY STRUCTURES

Structures considered primary are those which have formed during deposition of the sedimentary beds which contain them. Most of the structures described below are to be found in one or more of the outcrops which comprise the measured section near Hurley, Wisconsin (see STRATIGRAPHY AND LITHOLOGY). At those seven outcrops exposure is good and induration due to low grade metamorphism has helped preserve internal bedding features. Differential weathering on the outcrops has enhanced differences in relief and color between argillaceous and sandy rocks facilitating observation of primary sedimentary structures.

Description of External Bedding Features

The single most striking feature of the Tyler outcrops is the rhythmic interbedding of argillaceous and sandy beds (see Fig. 12). The dark brown to black-weathering argillites and slates are sandwiched between gray sandy beds. These units generally alternate one for one, but amalgamated sandy beds (Walker, 1967) are not uncommon. Sand beds are of two types, those which exhibit definite sequences of internal features and those which do not. The beds generally extend the length of the outcrop. Lower bedding contacts are generally very sharp while upper contacts are either sharp or gradational. The fine-grained rocks fracture parallel to bedding with the exception of the argillaceous units which break parallel to slaty cleavage where present.

Figure 12--U. S. HIGHWAY 2 ROAD CUT JUST OUTSIDE OF HURLEY, WISCONSIN SHOWING CHARACTERISTIC INTERBEDDING OF SLATES (OR ARGILLITES) AND GRAYWACKES. (SE 1/4, NE 1/4, S14, T46N, R2E, near Hurley, Wisconsin)

The coarser, sandy beds fracture haphazardly.

In general, the beds are very regular in appearance. Lensing of the sand units is not common but is present. Loading has disrupted the beds in some instances and undulatory bedding which is thought to be a primary feature (rather than tectonic or soft sediment deformation) is noted and described later. In one place, three sand beds separated by argillites merge into one when followed along strike. In other places fracturing and discontinuous argillaceous units test the integrity of a single sand unit along strike. Elsewhere two pinching sand beds between three argillites is suggestive of two possible explanations, localized erosion subsequent to deposition or true sand lenses. Which explanation is correct is not readily discernable in outcrop since the other end of the presumed lens is not exposed. The pinching of both the sandstones and argillites toward the exposed end seems to suggest lensing since erosion would abruptly truncate otherwise uniform bedding. There are other cases of unmistakable sand lenses. These are most evident where found within thick argillaceous sequences and only recognized with difficulty where found between other sand beds. Lenses vary greatly in size, from one centimeter thick beds to one meter thick beds. Most are small. Lensing, undulatory bedding and other bedding irregularities are rare.

Bedding thickness varies from very thin (less than 3 cm.) to very thick (more than 1 m.). Any single outcrop commonly exhibits all variations in bedding thickness. For this reason stratigraphic subdivision of the Tyler Formation cannot be based upon dominant bed thickness even though certain trends throughout the measured section are noted (see STRATIGRAPHY AND LITHOLOGY).

A total of seventeen-hundred and three beds were examined and measured in detail yielding a total thickness of 347.3 meters, or an average bed thickness of 20 cm. In general, the argillaceous beds tended to be thinner than the average, the sand units thicker.

Description of Internal Bedding Features

Internally, the sand beds often exhibit no primary sedimentary structures in outcrop. Such units are referred to as structureless beds and comprise 35% of the beds measured. It should be noted, however, that Hamblin (1962) and others have shown that examination of slabs by X-radiography, staining and etching will often reveal internal laminations and/or cross-laminations where the naked eye cannot. The beds in the Tyler Formation which are structureless in outcrop are referred to as sand or silt beds, according to grain size, and are distinguished from other sand beds which are characterized by specific combinations of internal structures (see below).

Laminations are common in some sand beds and in the upper silty and muddy portions of graded sand beds but are not evident in the interbedded argillites and slates. Table 8 summarizes various parameters measured in the laminated units.

Detailed studies have revealed that alternation of grain size, composition and mud content are the primary causes of laminations in other rocks (Middleton and Hampton, 1973). As can be seen from Table 8, laminations occur frequently in the Tyler Formation in fine sand and silt, but rarely in coarse and medium sand. Alternation of grain size, especially fine sand and silt or mud, and the accompanying change in composition (from dominantly quartz to dominantly micas and clays) is the cause of the laminations. Laminae in coarser sand units, on the other hand, are probably not the result of alternating grain sizes but rather are due to the presence within the sand unit of bands rich in muddy fragments (see PETROLOGY).

TABLE 8--THICKNESS AND GRAIN SIZE OF LAMINATED BEDS

Bouma Interval	Average Grain Size*	Average Thickness Of Bed (cm)	Thickness Of Laminae (cm)	Bouma Interval	Average Grain Size*	Average Thickness Of Bed (cm)	Thickness Of Laminae (cm)	Bouma Interval	Average Grain Size*	Average Thickness Of Bed (cm)	Thickness Of Laminae (cm)	Bouma Interval	Average Grain Size*	Average Thickness Of Bed (cm)	Thickness Of Laminae (cm)
B	s-m	.8	.3-t'	B	fs-m	13	.2-t	B	ms	4	1-t	B	m	4	t
D	s-m	4-2.5	.4-t	B	ms	1	.5-t	B	ms	15	t	B	fs	5.5	.2-t
D	m	3.5-.3	.3-t	B	ms	1.5	i	B	s		i	B	m	18	t
D	m	.2		B	ms	9	i	B	fs-s	4.5	.2-t	B	fs	4	t
D	m	3.8-2.5	1.5-t	B	ms	1.5	i	B	fs		t	B	ms	4.5	.2-t
D	m	3.5-2.7	1.3-.4	B	s	4	i	B	fs-s	10	t	B	s	2.5	.2-t
D	m	3-2.7	1-t	B	fs	5	i	B	fs-s	8.5	t	B	fs	2.5	t
D	m	2.4	t	B	s-m	5.5	.2	B	fs-s	30	1-t	B	fs	5.3	t
D	m	2.5	1-.1	B	s	2-1.5		B	fs	7	t	B	s	2.5	t
D	m	1.5	.5-t	B	m	10	5-t	B	m	6.5	t	B	m	3.5	t
D	m	4.5-3.1	1-t	B	s	4.5	i	B	fs	6.2	t	B	fs	3.5	t
B	s	2	.2	B	m	22	t	B	ms	10		B	fs	2.5	t
B	s	1		B	fs	.2	t	B	fs	41.5	1-t	B	s	14	t
B	s	3	1-t	B	s	9	i	B	m	13	.5	B	s	2	t
B	fs	6	1-.3	B	ms-fs	3.8	i	B	fs	6	t	B	s	1.5	t
B	s	4.5	1-.4	B	fs	.7	i	B	fs-s	1	.1	D	m	2	t
B	fs		t	B	fs	3.4-2.6	i	B	fs-s	1	.1	D	m	1-1.5	.9-t
B	fs	6	1	B	fs	1-.5	.2-t	B	fs	6	.5-t	D	m	3.5-4	1.2-t
B	ms-fs	8	.3-.4	B	fs-s	.8	.1-t	B	fs	4	.1	D	m	1	.4-t
B	fs	3	1-.5	B	ms	5.3	1-t	B	fs	23	.1	D	m	2.7-1.9	.5-t
B	ms	3	1-.1	B	fs-s	3.5	i	B	fs	3.5	i	D	m	3.5-3.1	.3-t
B	fs	1	.1-t	B	fs-s	12	i	B	fs	12.9	1-t	D	m	1.4-1.2	.3-t
B	m	2.2	.5-1	B	fs-s	9	.3-t	B	fs-s	21	t	D	m	1	.3-t
B	m		.3	B	fs	15	1-t	B	fs	3	1-t	D	m	1.5	.3-t
B	fs	5	.2	B	fs-s	5	i	B		3-4	t	D	m	.3	.1-t
B	fs		.2-t	B	fs-s	7	1-t	B	fs	3	1-t	D	m	.7-.2	t
B	s	2	.3	B	fs-s	8.2	t	B	fs	13.7	i	D	m	.8	.2-t
B	cs-ms	33.9	6-t	B	s	1	i	B	fs-s		i	D	m		.5-t
B	m	5.5-5.1	.5-.1	B	s	1.5	i	B	m	2	t	D	m	1.5	.4-t
B	s-m	1.5	.1-.3	B	fs	6	3-t	B	ms	6	i	D	m	2	.6-t
B	s	1	.1-.4	B	fs	6.6	.7-t	B	m	11		D	s-m	3	1
B	fs-s	5	i"	B	fs-s	13		B	fs	.3	t	D	m	1.5	.5-t
B	m	7	.4-t	B	fs	7	t	B	fs-s	3	i	D	s-m	.5	
B	fs-s	2.3	.7-t	B	fs	5.2	1-t	B	fs	2	i	D	s-m	7.5	.5-t
B	s	2	.4-1	B	ms	.7	i	B	fs	4	.5-t	D	m	.3-4	.3-.2
B	s	2	.7-t	B	ms	12	1-.2	B	s	.5	t	D	m	2	.2-t
B	fs-s	5	1-t	B	ms	12	t	B	fs-s	1.4	t	D		1.4	t
B	m	2.5	.1-t	B	fs-m	5.5	.3-t	B	m	4.5	t	D	s	10.5	t
B	s	10	.2	B	m	5	1-t	B	s	1	t	D	m	15	t
B	s	15	.5-t	B	ms	.7	.3-t	B	fs	.5	t	D	ms	3.5	.2-t
B	s	2-6	.2-t	B	fs	.4	t	B	fs	13.5	t	D	fs	9	t
B	ms	6	.2	B	m	2.3	t	B	fs-s	12	t	D	fs-m	2	.5-t
B	ms	6	i	B	fs-s	1	t	B	fs	7	t	D	s	.5	i
B	ms-fs	2	.1	B	ms	13	1	B	fs	1	t	D	s	2	.1-.3
B	fs-s	4	t	B	ms	2	i	B	m	5	t	D	s-m	15	.2-t

*m=mud, s=silt, fs=fine sand, ms=medium sand, cs=coarse sand
't=tissue thin
"i=indistinct

Thirteen percent of the 1002 sand-silt beds measured in detail exhibit graded bedding in outcrop. Middleton (1967) has defined two types of grading by experimental studies. Coarse tail grading is grading only within the coarsest percentile of grain size and distribution grading is defined as grading of the entire size distribution. Coarse tail grading is not recognized in the rocks of the Tyler Formation. Several variations of distribution grading were recorded including normal distribution grading, multiple normal, reverse and multiple reverse.

The great majority (90%) of the graded beds are normally graded. These beds grade from granules, coarse, medium or fine sand at the base to medium, fine sand or silt at the top. Only 28 beds grade upward into mud. The presence of lutite clasts of pebble size or larger at the base of the unit often enhances the graded appearance of the bed. Just as great grain size variation can enhance graded appearance, so also will lack of coarse particles obscure grading. Often the beds are so fine grained that grading, if present, is not discernable. Thickness of graded beds ranges from very thin (less than 1 cm.) to very thick (more than 2 m.). In the thick beds grading is hard to see since the change in average grain size is very gradual from bottom to top. Some graded units are laminated throughout.

A dozen beds exhibit multiple or recurrent grading (Kuenen, 1953) where the graded unit repeats itself as many as five times without an intervening pelagic bed (see Fig. 13). Sometimes the graded units

Figure 13--MULTIPLE GRADED BED. Note three-fold repetition of graded unit. Pencil points up section. (same location as Figure 12)

within these multiples are separated by faint laminations and/or zones rich in mud clasts which suggest the disturbance of slightly older pelagic beds. The multiple graded beds vary in thickness from 4.5 cm. to 221 cm.

Two reversely graded beds were found, one of them showing multiple reverse grading. The graded intervals of this multiple reverse bed are separated by lutite clast-rich zones as were some of the multiple normal graded beds.

Apparent reverse grading similar to that described in the McKim Formation of the Elliot Lake Group by Robertson and Card (1972) was noted in a few outcrops (Fig. 14). The apparent grading is the result of the development of metamorphic minerals in the upper muddy parts of normally graded beds which have grown larger than the sand grains at the bottoms of the beds. Pitting on the beds due to preferential weathering of these porphyroblasts is common. The pits become larger and more numerous as the size and density of the metamorphic minerals increase. In a graded bed this gives the appearance of grading opposite the true grain size gradient. Pitting is especially common in the western part of the study area where slates with large porphyroblasts have been called "spotted slates" (see PETROLOGY).

Microcross-bedding of the trough type is common in the sand and silt beds of the Tyler Formation (Fig. 15). The average amplitude of cross-sets is 1.14 cm. with a range from .4 - 4 cm. The average wave length (where measurable) is 7.67 cm. with a range from 3 - 15 cm. The cross-laminae are invariably truncated at the top by the overlying sediments so true average amplitude is undoubtedly greater than that cited.

Cross-bedding occurs in medium sand, fine sand and silt but is never found in coarse sand. The average thickness of beds with cross-laminae is 25.2 cm. with a range from 2 to 88.5 cm. The cross-laminae, which do not encompass the entire bed thickness, are visible because of

Figure 14--APPARENT REVERSE GRADING IN THE TYLER FORMATION. True tops lie in the direction of the arrow. The rough, "coarse" parts of the beds (A) are really the muddiest (and finest-grained) while the smoother areas (B) are composed of sand and silt-sized grains. The rough texture of the muddy areas is caused by the preferential weathering of metamorphic porphyroblasts. Scale is in centimeters. (NE 1/4, NW 1/4, S12, T44N, R3W, near Mellen, Wisconsin)

mud concentration along the foreset beds.

Mud lenses found in conjunction with cross-laminae near the tops of sand beds are strongly reminiscent of flaser bedding which is commonly found in recent tidal flat sediments. The appearance of these

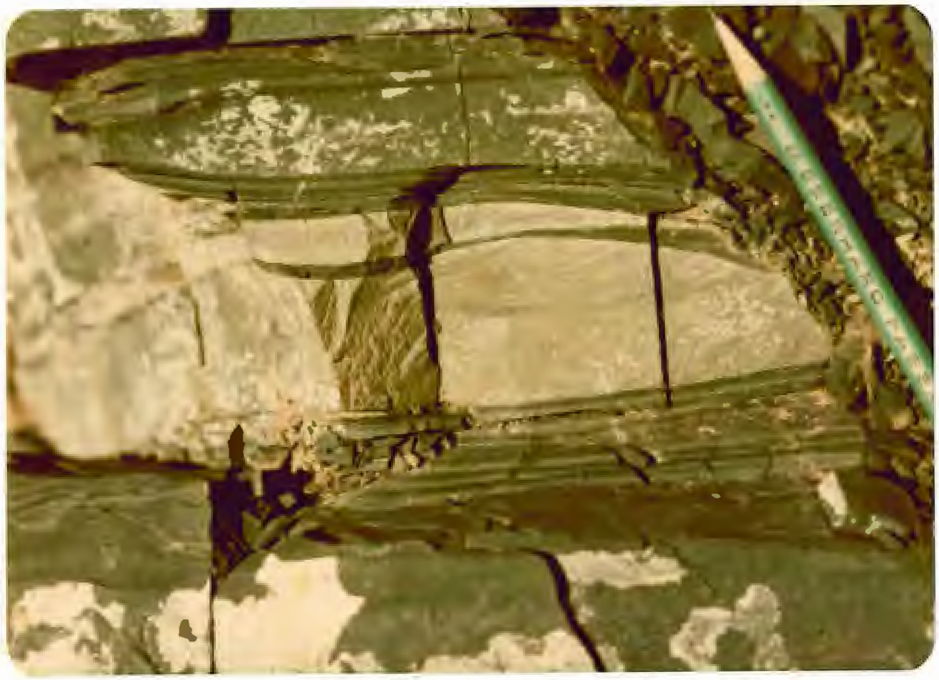


Figure 15--MICROCROSS-BEDDING IN A FINE GRAINED SANDSTONE BED. (SE 1/4, NE 1/4, S14, T46N, R2E, near Hurley, Wisconsin)

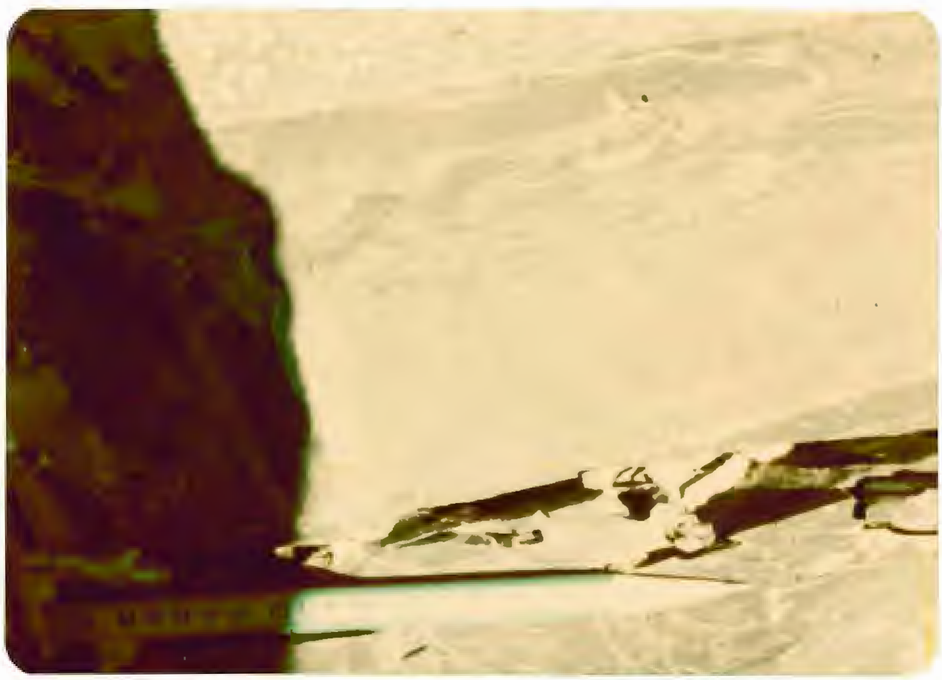


Figure 16--FLASER BEDDING (?) IN A GRAYWACKE. Note the lense-like appearance of mud (dark colored) apparently trapped in troughs between ripple crests. Note also the grading of the graywacke and the loading and subsequent deformation of the overlying thin argillite. (Same location as Figure 15.)

mud lenses in profile (see Fig. 16) suggests the entrapment of mud in troughs between ripples with sinuous crests (e.g. Blatt, Middleton and Murray, page 156).

Linguoid ripple marks are recognized with difficulty on the surfaces of some bedding planes (Fig. 17). These ripple marks were distinguished from soft sediment deformation features by the regularity of troughs and crests. Loaded bedding surfaces tend to be very irregular (Fig. 18) and don't exhibit consistent intervals between crest and trough as rippled surfaces do (Fig. 17). Twelve bedding surfaces with linguoid ripple marks were found in the Tyler. All were found in the central

Figure 17--LINGUOID (?) RIPPLE MARKS ON THE TOP OF A GRAYWACKE UNIT. The surface of the graywacke is covered by a thin veneer of argillite. (NE 1/4, SW 1/4, S29, T46N, R2E, near Montreal, Wisconsin)

Figure 18--LOADED SOLE OF A GRAY-WACKE BED. Note the irregularity of the loading. (just off the County Line Road between Ashland and Iron Counties, Wisconsin; NE 1/4, NE 1/4, S36, T45N, R2W)

beds of the measured section and in close proximity (stratigraphically and geographically) to undulatory bedding surfaces described below.

Table 9 shows that there is no apparent relationship between bed thickness and the occurrence of ripple marks. Also note that linguoid ripple marks occur in otherwise structureless bedding as well as in association with other internal bedding features, especially laminations. Straight-crested (or moderately sinuous) ripples were found on three

TABLE 9--BED THICKNESS AND GRAIN SIZE OF BEDS WITH RIPPLE MARKS

Type of ripple marks	Grain size	Bed thickness (cm)
Linguoid.....	fine sand	5.5
	medium sand	3.5
	medium sand	10
	medium-coarse sand	54
	medium-coarse sand	32.5
	medium-coarse sand	17
	fine-medium sand	7
	fine sand	6.5
	fine sand	12
	medium sand	11
	?	59
	medium sand	11
Straight-crested..	medium sand	41
	coarse sand	71.5
	medium sand	41.5
Ripples seen in profile only....	medium sand	22.5
	fine-medium sand	4.5

beds which contained internal laminae. Where bedding planes were not exposed, ripple marks were recognized in profile by wavy bedding and by the localization of muddy sediments in the troughs of the presumed ripples.

Miscellaneous internal bedding features include lutite clasts, concentration of large clastic grains in loaded scour marks, large scale undulatory bedding resembling standing sand waves or sand ridges (e.g. Blatt, Middleton and Murray, 1973) and large scour marks.

The lutite clasts vary greatly in size and concentration within both the turbidites and the structureless sand and silt beds. The largest clast measured 25 x 61 cm. (in two dimensions). The smallest was microscopic. Often they are so numerous in a sand bed that the unit looks conglomerate (Fig. 19). The lutite clasts may be concentrated at any level in a sand bed or dispersed randomly. Often only one or two clasts are seen "floating" in sand. Rarely a sand bed contains a string of clasts laid end to end. Two or more lutite clast-rich zones in a single sand bed is not uncommon.

The clasts are most often angular, elongated and truncated by a series of step-like breaks. Sometimes they appear "wispy" as if stretched before breaking. Others are paper thin and very delicate-looking. These latter, "toe-nail-shaped" clasts, look much like the lenticular muds of flaser bedding but are not found in association with cross-laminae. Fig. 20 shows folding within a laminated lutite clast suggesting that it was still soft when caught up in the enclosing sand.

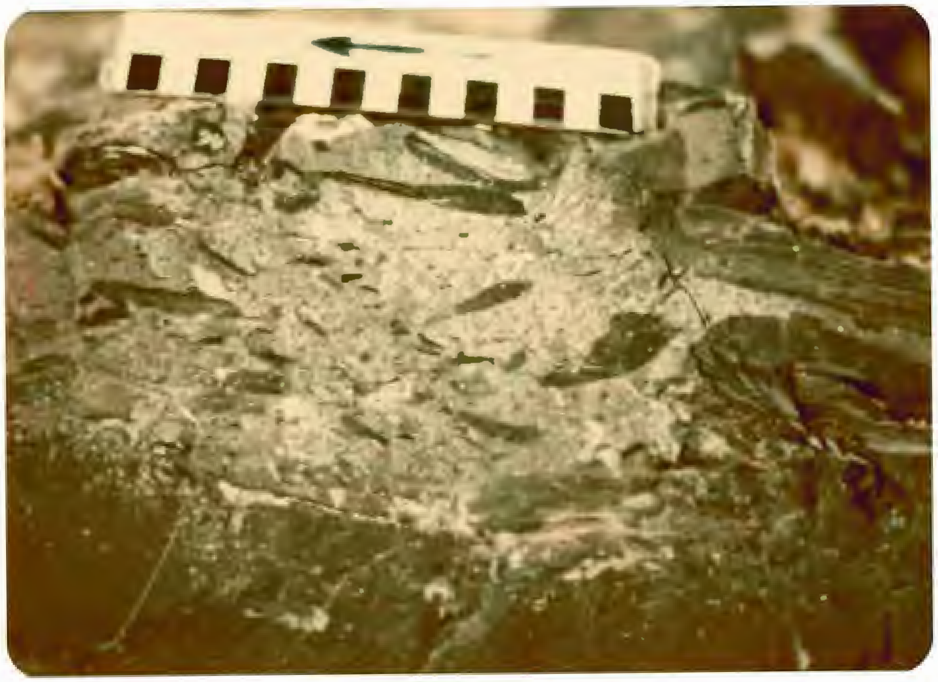


Figure 19--INTRAFORMATIONAL MUD-CHIP CONGLOMERATE. Scale is in centimeters. (SE 1/4, NE 1/4, S14, T46N, R2E, near Hurley, Wisconsin)

Argillaceous clasts are sometimes found in close association with other features which indicate relatively strong currents, like scour marks filled with coarse sand. Lutite clasts are common to both structureless beds and beds with other internal features. Rarely, imbricated clasts suggest current orientation which is in agreement with other paleocurrent indicators.

Loading of soft sediment is a secondary phenomena, but the loci of the loading pockets was probably controlled by bedding irregularities due to primary sedimentary structures. Figure 21 shows a concentration of coarse sand grains in loaded pockets which suggests a primary depression in the bedding plane. This depression, probably a scour mark, caught the coarse grains as the overlying bed was being deposited.

Figure 20--FOLDED LUTITE CLAST WHICH WAS CAUGHT UP IN THE CURRENT WHICH DEPOSITED THE ENCLOSING GRAYWACKE BED. (same location as Figure 19)

During compaction, density difference between the sand above and the mud below accentuated the sand bulge and created the small flame structure shown. More convincing scour marks are noted elsewhere (see below and see Paleocurrent Indicators).

The features referred to as standing waves or sand ridges show relief on the order of 10 to 20 cm. (Fig. 22). A tectonic origin of those features doesn't seem likely. The undulatory bedding planes which

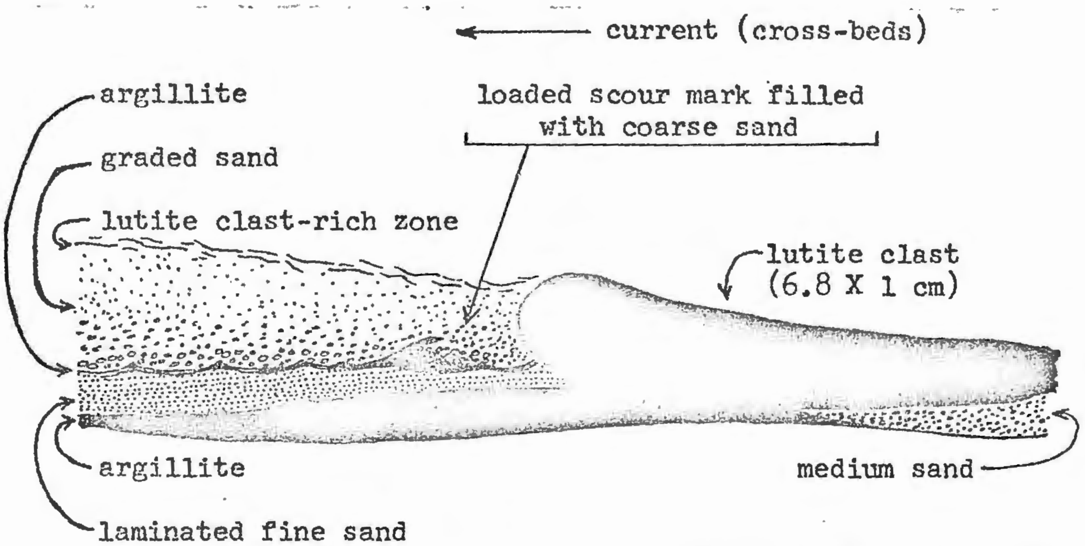


Figure 21--SKETCH OF PRIMARY SEDIMENTARY FEATURES SEEN IN A GRAYWACKE BED. (same location as Figure 19)

characterize the presumed waves or ridges are accompanied by large scour marks (Fig. 23) and are quite numerous locally but not persistent throughout the outcrop. If these features are tectonic, why are some sand beds affected and not others? It is believed that they are primary sedimentary structures. There is little doubt that the large scour marks which accompany them are primary. Internal primary sedimentary structures, with the exception of rare laminations, are not found in argillaceous beds.

Sequences of Internal Bedding Features

Internal bedding features often have a definite sequence identical to that found in other formations (Fig. 24). A general sequence was first described by Bouma (1962). Figure 25 pictures the ideal Bouma sequence and the variations on the ideal that are found in the Tyler

Figure 22--UNDULATORY BEDDING SURFACE (SAND RIDGE?) OF THICK GRAYWACKE BED. Note hammer for scale. (NE 1/4, SE 1/4, S29, T46N, R2E, near Montreal, Wisconsin)

Figure 23--LARGE SCALE SCOUR MARKS ON TOP OF A THICK GRAYWACKE BED. The bottom of the bed shown in Figure 22 can be seen in the upper right corner of of this photo. Current movement from upper left to lower right, parallel to hammer handle. (same location as Figure 22)

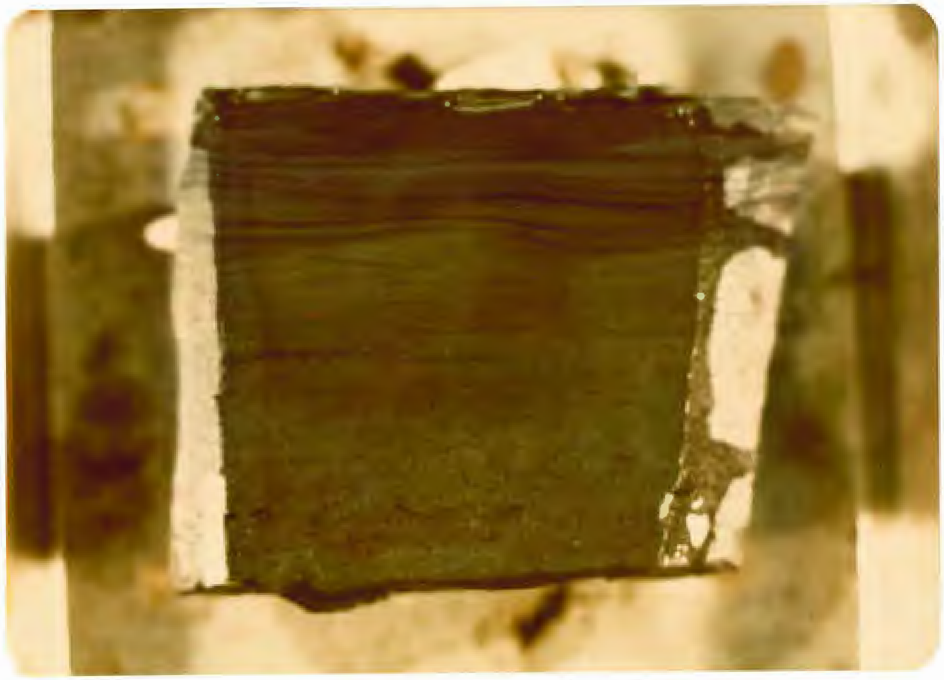


Figure 24--SLAB OF TYLER GRAYWACKE BED SHOWING SEQUENCE OF INTERNAL BEDDING FEATURES. From bottom to top; a graded unit, a lower laminated unit, a cross-bedded unit, an upper laminated unit and a mud unit. Slab is approximately 12 cm. wide from top to bottom. (SE 1/4, NE 1/4, S14, T46N, R2E, Hurley, Wisconsin)

Formation. In the ideal Bouma bed the lowermost part of the bed is graded (T_a) and is followed successively by a lower laminated unit (T_b), a cross-bedded unit (T_c), an upper laminated unit (T_d) and a mud unit (T_e).

It was noted while preparing Figure 25 that not only were there definite bedding sequences within sand and silt units, but the sequences themselves occur in groups. That is, several T_a beds occur in succession, then several T_{ab} beds, then several T_b beds, etc. This phenomenon is illustrated by the T_b (Bouma's lower laminated division) beds in Fig. 26. The Bouma T_b division is relatively rare throughout

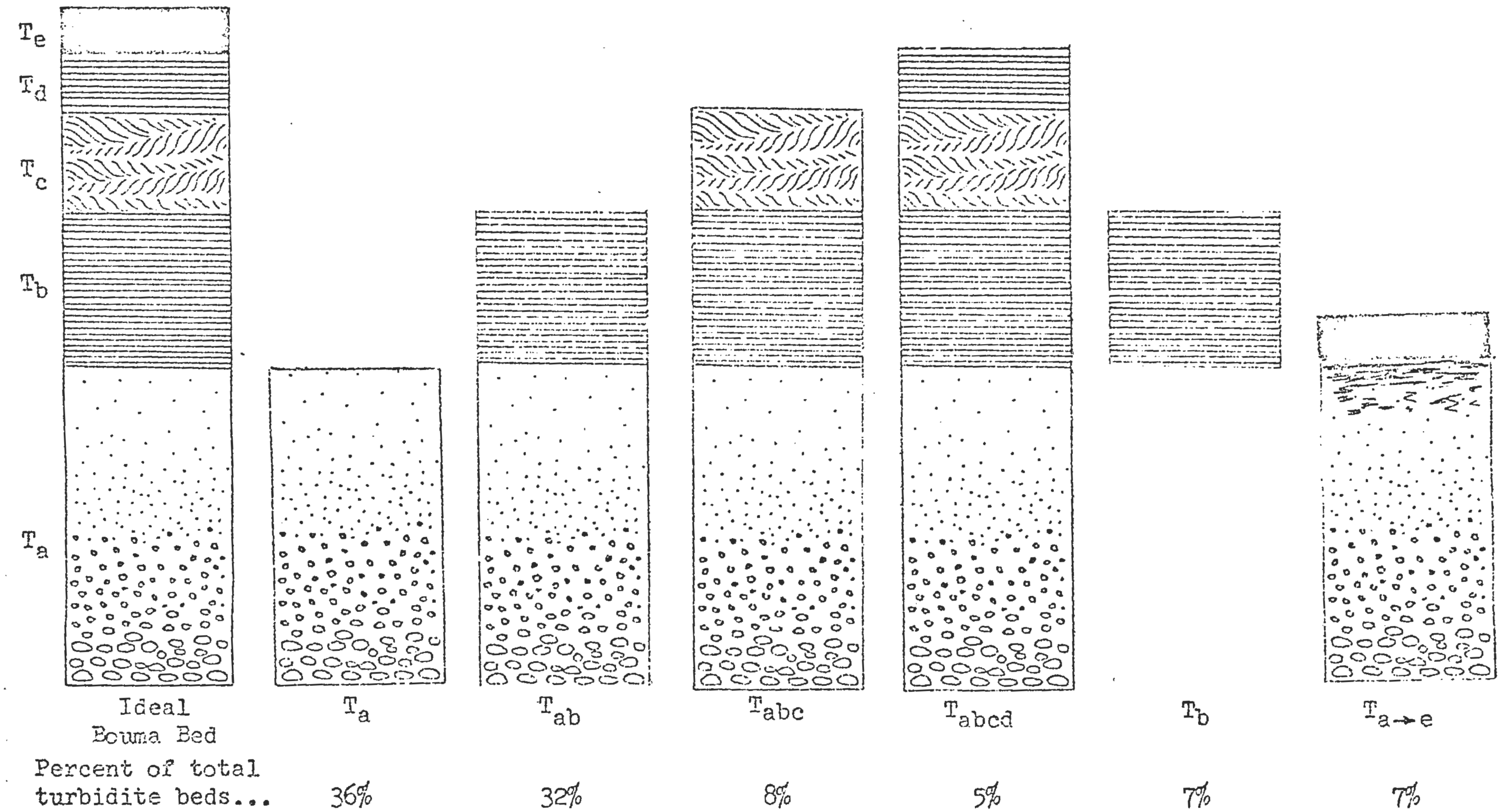


Figure 25--THE IDEAL BOUMA BED AND VARIATIONS ON THE IDEAL SEEN IN THE TYLER FORMATION. Total of 412 beds. (after Morey and Ojakangas, 1970)

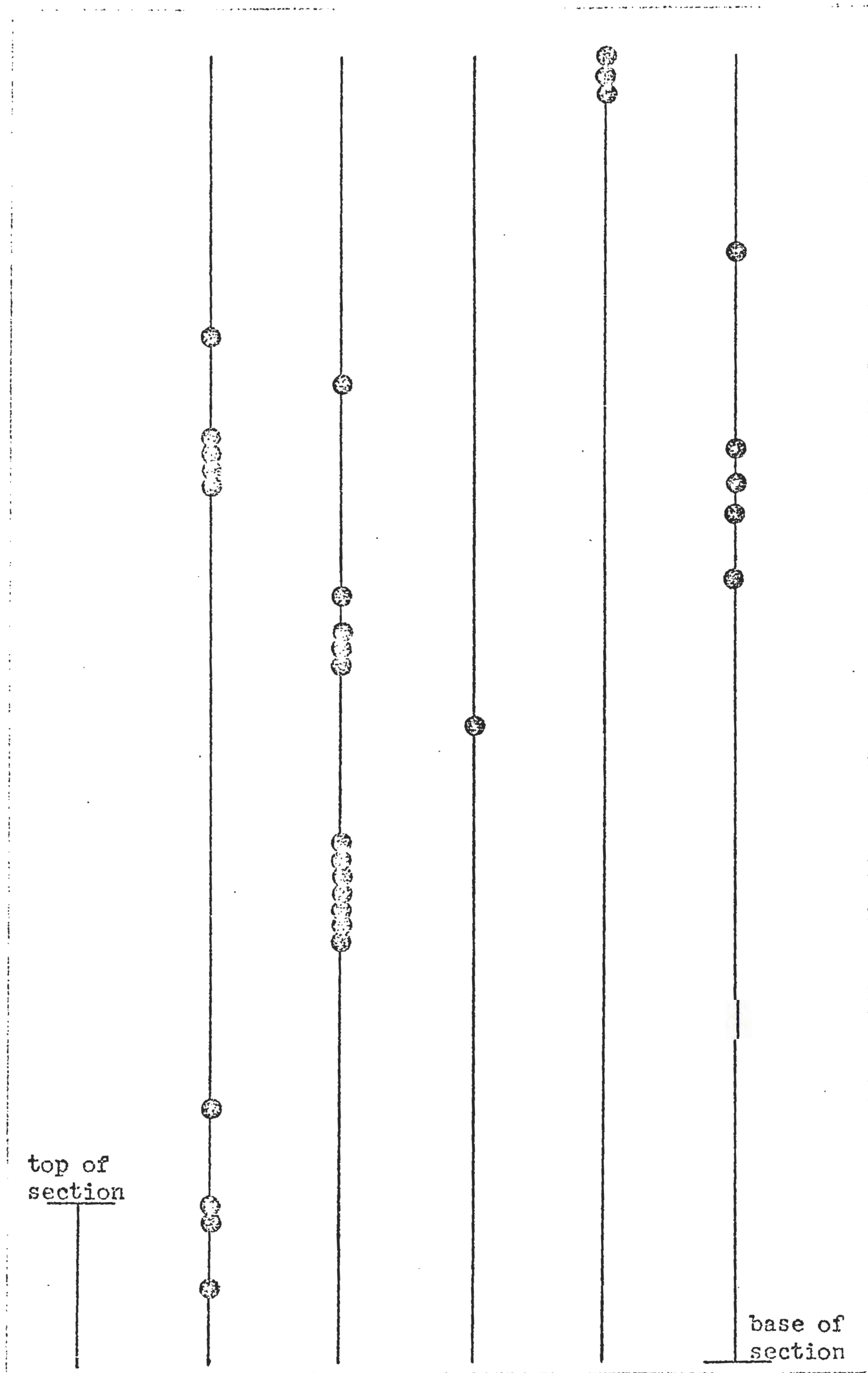


Figure 26--STRATIGRAPHIC GROUPING OF BOUMA T_b BEDS IN THE MEASURED SECTION. Each dot represents one T_b bed. Each blank space of width equal to one dot represents a graywacke bed of different internal character. Argillaceous beds are not included.

the measured section as a distinct bed (although in combination with other Bouma intervals it is quite common). Yet when it does occur individually, it occurs three or four times in a row. Toward the top of the section it occurs seven times without an intervening sand bed of different internal character. The other Bouma subdivisions seem to exhibit the same sequential occurrence, although it is not as clear as the T_b units.

Discussion and Implications of Bedding Features

Rhythmic interbedding of fine grained sediments and coarser-grained sediments, normally not found in the same depositional environment, requires explanation. It is not believed that the depocenter fluctuated so rapidly as to allow for the deposition of both sediments in their respective environments, but rather that the pelagic units were deposited under deep water conditions which were periodically disturbed by abnormal events. These abnormal events, which are believed to have been turbidity currents, were responsible for deposition of the anomalous sands.

Sharp top and bottom bedding contacts suggest that there was no gradual change from one environment to the other, but rather that the quiet water pelagic environment was disturbed by geologically instantaneous events. Amalgamated sand beds suggest that these turbidity current events occurred periodically with such rapidity that intervening normal sediments (pelagics) were not deposited or, alternatively, that the events were possessed of sufficient force to completely remove the previously deposited mud bed. Amalgamated sand beds could also reflect several pulses within a single depositional event.

Regularity of bedding thickness along strike (with very few lenses, bifurcating beds, etc.) suggests that these events were laterally extensive. Very regular bedding is characteristic of the Tyler Formation as a whole, indicating that the same kinds of processes were operating throughout the depocenter.

Blatt, Middleton and Murray (1972) state that flaser bedding (lenticular mud laminae trapped between ripple crests) is most commonly found in tidal flat sediments but may also occur in deltaic, lacustrine and sheet-flood deposits. The environmental requirements are alternating periods of active currents and quiescence (Reineck and Wanderlick, 1968) with the sand ripples being formed when currents were active and mud deposited during slack water time. The mud is caught in ripple troughs. If mud was deposited on the crests it is removed by the subsequent current which builds another rippled layer. In either case, the result is lenticular mud laminae in the old troughs. Study of the Tyler Formation suggests that flaser bedding can also occur in deep water environments. Alternating periods of movement and quiescence are found in such environments and can occur within a single depositional event as shown by flaser lenses found within a single bed. Depositional mechanisms are discussed below.

The presence of lutite clasts suggests that the water which deposited the sands wherein they lie was moving at a velocity much greater than that which would allow the settling of clay-sized sediments. The morphology of the clasts suggests that these were pieces of a mud bed ripped up by a strong current. The site of deposition of the sands which contain them must have been far enough out in the basin of deposition so that the current passed over pelagic sediments on its way to the depocenter. That is, the current passed through an environment which lay below wave base on its way to deeper water. The wispy and stretched shapes of some clasts suggest a cohesive mud which yielded to erosion only under considerable persuasion from the passing current. One of the clasts (Fig. 20) shows internal folding which indicates entrainment before lithification.

The concentration of lutite clasts at the bottoms of some beds and in zones throughout other beds can be interpreted in two ways. It seems likely that clasts concentrated in basal zones of graded beds were part of the sediment load of the same current which deposited the sand and their presence is the result of the same fluctuating hydraulic conditions which caused the graded bedding. That is, they are clasts brought to the depo-site by currents. Their positions within the resulting sand bed are dependent upon their settling velocities relative to the settling velocities of the surrounding sediments. Alternatively, some beds contain very numerous mud clasts with no consistent size variation from bottom to top, suggesting they may be locally derived and may very nearly be in situ clasts, having moved only a short distance. Such clasts could be the remnants of a disturbed mud bed. The large size of some of the clasts suggests a very competent current, probably with internal turbulence. Other features which suggest turbulent flow are discussed below and in Appendix C in which theoretical studies are applied to observations in the Tyler Formation. By utilizing clast size to give some idea of current competence, the velocity of a specific turbidity current is calculated at 44 cm $sec.^{-1}$ (Appendix C).

As mentioned earlier, a few sand beds contain stringers of aligned lutite clasts. These stringers are hard to explain in terms of a strong turbulent current which other features seem to suggest. It is unlikely that clasts of different sizes and shapes (and therefore of different hydraulic character) would settle out of a waning current at the same time and be deposited at the same horizon within the sand bed. On the other hand, it is possible that these stringers are, like some of the basal mud-chip conglomerates described above, merely remnants of a once continuous mud unit. But if that is the case, why should the current pick up some clasts and leave behind others of differing shapes and sizes? Also, why should a turbulent current leave behind a layer exactly one clast thick? And, if clast stringers are in situ remnants of bedding, then the sand bed above and the sand bed below are the result of two different events, perhaps separated by many years. They should show some differences in texture and/or composition, however slight, if this were the case. Examination of these beds shows no such differences. These clast stringers are thought to be indicative of grain-flow origin and are discussed in more detail later.

Loaded pockets with coarse sand in them (Fig. 21) suggest loaded scour marks. Scour marks are the result of turbulent eddies rotating about near horizontal axes transverse to current flow which locally erode bedding surfaces (Allen, 1971a). Again, the evidence suggests a turbulent current strong enough to erode cohesive muds and strong enough to carry coarse sediment and large clasts.

The undulatory bedding surfaces (Fig. 22) referred to as standing sand waves or sand ridges provide more evidence for strong currents which periodically disturbed an otherwise calm depocenter. It is unlikely that these features are standing sand waves. Apart from the difficulty of preserving them (the waves would tend to be destroyed or modified into alternative bed forms as regime dropped) there is evidence that the flow was not laminar as is required for standing sand waves. The large scale scour marks found on one of these beds (Fig. 23) suggests turbulent flow (Allen, 1971a).

It is also important to note that the crests of these bedding undulations are parallel to inferred current direction, not perpendicular as crests of standing sand waves should be. Current direction has been ascertained not only from indicators in the surrounding beds, but also from the scour marks found on the undulatory beds themselves. Thus, it seems unlikely that these are standing sand waves.

The more reasonable explanation is that the undulatory beds are subaqueous sand ridges similar to tidal sand ridges. Tidal sand ridges are thought to be the result of secondary, spiraling currents acting within a larger current (Blatt, Middleton and Murray, 1972; Houbolt, 1968; Harms, et al., 1975). These helicoidal currents pile sand in ridges parallel to the major flow direction. The hydrodynamics of sand ridge formation are not well understood, but it seems that the ridges are created in areas characterized by laterally extensive (as opposed to channelized) currents. It is suggested that such currents were operating in the Tyler depocenter. It is further suggested that the velocities of the currents which deposited and/or built these sand ridges were the highest current velocities operating in the Tyler basin, for all other bed forms suggest currents of lesser magnitude.

The sequences of internal bedding features, the Bouma sequences are indicative of turbidity current deposition. Harms and Fahnestock (1965), Walker (1965), Walton (1967), Middleton (1967) and others have interpreted the Bouma sequence in terms of the flow regime concept which suggests deposition of the sediments by a waning current. The current was laden with sediment of varying grain sizes. The decreasing competency of the waning current allowed differential settling of sizes (grading) and creation of various traction (cross-bedding, ripplemarks) and non-traction (laminations) features. The Bouma sequence documents decreasing velocity up-sequence, increasing importance of traction up-sequence and decreasing flow regime up-sequence.

It has been shown that sediment-laden current like that under consideration flow because of the sediment load which imparts to the water a higher effective density than the surrounding, cleaner ambient water. The only prerequisite for flow is a mechanism which will put the sediments into suspension (see below). Once suspension is achieved, flow is initiated and perpetuated by the force of gravity acting on the more dense fluid. Such currents are called density currents or turbidity currents because of the internal turbulence necessary to support the sediments. It is believed that the sandstones and siltstones of the Tyler Formation which exhibit the features described above (especially the Bouma sequence) were deposited by turbidity currents. Further evidence suggesting turbidity current deposition can be found in the section on Paleocurrent Indicators.

The author is unable to satisfactorily explain the groups of Bouma sequences depicted in Figure 26. Perhaps they have something to do with the proximality of the depo-site to the place of initiation of the current (as per which of the several Bouma intervals were created). This line of reasoning would seem to suggest several different geographic sites of current initiation at varying distances from the depo-site. For reasons unknown, several currents were initiated at one site during a certain span of time followed by several currents from a different site, etc. If these separate sites existed contemporaneously, then paleocurrent indicators should indicate different flow directions for currents from the different sites. Such is not the case (see Paleocurrent Indicators).

Alternative, and perhaps more reasonable explanations include: periodic displacement of the initiation site by tectonic movement in the source area; varying intensity (competency) of successive flows from a single initiation site; varying amounts of sediment and varying grain sizes in sediment supplied to the initiation site (perhaps due to climatic changes or more local weather phenomena); or investigator error due to non observation of internal structures obscured by metamorphism and/or diagenesis. Another alternative which will be developed further in this paper, is the lateral migration of the currents which deposited the rocks. The loci of those currents were the result of channel position.

Truncated Bouma sequences with the upper one, two or three intervals missing are most common in the Tyler Formation (see Figure 25). Interpretation of this observation in terms of proximity to source (Walker, 1967) suggests a very proximal site of deposition. However, studies of recent ocean bottom turbidite sediments and the morphologies of ocean bottom features (continental slope canyons, canyon fans, abyssal plains, etc.) suggest that simple calculation of the Walker proximality index may lead to erroneous conclusions. Proximal and distal environments are complicated by many factors. For example, channel overbank deposits on a subaqueous fan can build "distal turbidite sequences" (characterized by Bouma T_c, T_d, T_e, intervals) adjacent to coarse grained, conglomeratic deposits at the base of the continental slope. Also, ocean bottom currents flowing parallel to and on continental slopes can rework previously deposited sediments creating primary sedimentary structures such as small scale cross-beds and laminations which are strongly suggestive of distal turbidites (Bouma and Hollister, 1973). Such factors as these greatly complicate proximal-distal interpretation. Therefore, sequences of lithology and facies associations are probably more reliable indications of depo-site than the presence or absence of Bouma intervals.

It is hard to explain the structureless beds in terms of turbidity current deposition, since they do not exhibit the structures characteristic of turbidites. Although it has been suggested that Bouma T_a intervals can be massive and exhibit no internal grading (Bouma, 1962), it was decided to avoid naming these beds T_a units because that would imply deposition by turbidity currents, which cannot be proven. Likewise, structureless beds cannot be explained by traction current mechanisms (grain-by-grain deposition) because they lack traction current features (cross-bedding especially) and because many of them contain lutite clasts much larger than the surrounding sand grains. Rapid, chaotic submarine slumps can be ruled out by a lack of internal convolutions and, rarely, by the presence of lutite clast stringers which would certainly have been destroyed if slumping were the depositional mechanism.

The structureless beds resemble the kinds of sedimentary beds characteristic of grain-flow deposits. Grain-flows were first proposed by Bagnold (1954, 1956) on theoretical grounds. An ideal grain-flow bed was drawn by Stauffer (1967). That ideal bed is reproduced in Figure 27.

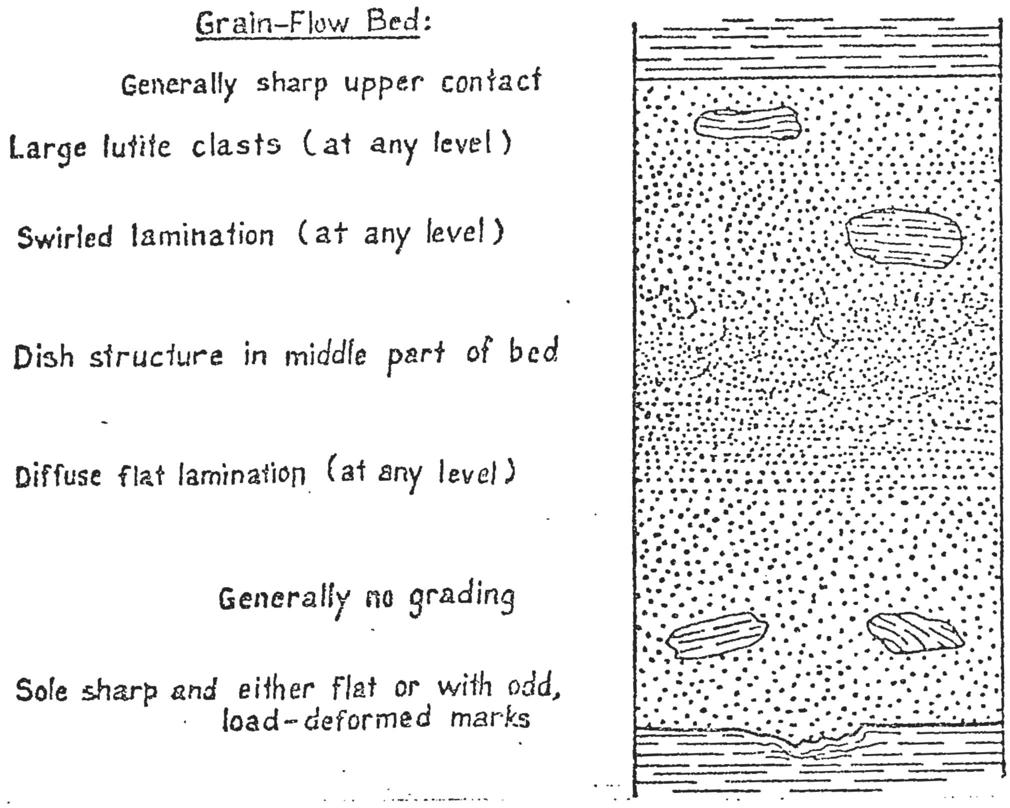


Figure 27--THE IDEAL GRAIN-FLOW BED (from Stauffer, 1967).

The grain-flow bed is characterized by sharp upper and lower contacts with miscellaneous sole marks sometimes developed at the bottom contact, large lutite clasts at any level in the bed, swirled and flat laminations at any level, dish structure (concave upward laminations) and lack of grading. On these grounds it is relatively easy to tell an ideal grain-flow bed from an ideal turbidite. But the beds in the Tyler formation are not ideal.

Consider the features characteristic of the structureless beds in the Tyler. They generally exhibit sharp lower contacts (characteristic of both turbidites and grain-flow beds), sharp upper contacts (characteristic of both grain-flow beds and truncated turbidites) and lack of grading and laminations (common in both incomplete turbidites and grain-flow beds). Some structureless beds contain very large lutite clasts which would undoubtedly qualify as "outsize" (Stauffer, 1967). But the majority of the lutite clasts probably are not "outsize." Dish structure and swirled laminations are absent. However, Stauffer observed dish structure only in beds greater than 4 meters thick. Such thick beds are rare in the Tyler Formation. There seems to be no positive proof of grain-flow deposition or turbidity current deposition in the structureless beds.

Theoretical considerations, however, suggest a grain-flow mechanism. Flow in turbidity currents is turbulent. Flow in grain-flows is laminar (Stauffer, 1967; Bagnold, 1954, 1956). It is suggested that the single layer clast stringers described above are good evidence of laminar flow. The clasts are introduced into the flow at some particular point upstream. (The erosive power of a laminar flow is open to some question. Documentation of the erosive capabilities of grain-flows is found in Shepard, 1961, Dill, 1966, and Shepard and Dill, 1966). The location of their introduction determined their position in the moving sand mass and, once placed there, stayed at that horizon even until deposition because of lack of turbulence.

Another line of reasoning involves the occurrence of reversely graded beds. Inverse grading, according to theories presented by Bagnold (1954) and Middleton (1970), should be evident in grain-flow deposits. Both examples of reverse grading found in the Tyler

Formation occur in beds lacking other internal sedimentary structures. They could be grain-flow deposits. However, reverse grading in two beds cannot be construed to be significant evidence of grain-flow origin for the other 588 structureless beds measured in the Tyler Formation. In fact, lack of numerous examples of reverse grading could be considered negative evidence.

Middleton (1970) has shown that slopes of 18 - 37° are necessary to maintain grain-flow movement wherein excessive pore pressure is absent. This is a significant problem when dealing with natural systems as slopes of that magnitude are rare. Those same studies indicated, however, that if the interstitial fluid is water, laden with mud (density = 2 gm. $cm.^{-3}$) then slope angle necessary for grain-flow is reduced to 4.5° - 10.5°. In the Tyler Formation, primary slope can be estimated from measurement of various primary sedimentary structures. Those calculations (Appendix D) yield a primary slope angle of $0^{\circ}10'$.

Although the author is strongly biased toward grain-flow origin of at least some of the structureless beds, it cannot be proven beyond citing the evidence already presented. Sediment transport probably exhibits both laminar and turbulent flow at some point in its journey to the depocenter. It is, therefore, probably best to think of grain-flows and turbidity currents as two members of a gradational spectrum of transport mechanisms as is suggested by Middleton and Hampton (1973). The kind of environment which fosters one could also foster the other.

Sanders (1965) suggests that grain-flows and turbidity currents are two separate phases of the same sediment flow event.

Documentation of earthquakes as initiating mechanisms for turbidity currents can be found in the literature. Overloading of sediments on deltas and at the heads of subaqueous continental slope canyons can also result in instantaneous release of great amounts of sediments which induce turbidity current flow. Both mechanisms could also cause grain-flows. The initiating mechanisms of the turbidity flows and grain-flows which deposited the Tyler sediments can never be known, but it is suggested that earthquakes which undoubtedly accompanied contemporaneous volcanism just east of the study area where the Copps Formation and the Presque Isle volcanics are interbedded (see GEOLOGIC SETTING) probably were a major contributing factor.

Under the heading STRATIGRAPHY AND LITHOLOGY a suggestion was made that some very thick sandstones at the top of the measured section were channel fill deposits. Now that bedding features have been described, further explanation should be made. Examinations of ancient and modern resedimented facies has led to the proposed facies classification system by Middleton and Bouma (1973) which is reproduced in Table 10. The coarse-grained rocks of the Tyler Formation fit well in that classification and are assigned to Facies B2 and C. Interbedded argillites and slates are Facies G.

A Facies Association chart (Figure 28) from the same publication (Middleton and Bouma, 1973) suggests that the association of facies B2

TABLE 10--BASIC CLASSIFICATION OF TURBIDITE AND OTHER RESEDIMENTED FACIES (from Middleton and Bouma, 1973)

BOUMA SEQUENCE NOT APPLICABLE	FACIES A -- Coarse grained sandstones and conglomerates A1 Disorganized conglomerates A2 Organized conglomerates A3 Disorganized pebbly sandstones A4 Organized pebbly sandstones. FACIES B -- Medium-fine to coarse sandstones B1 Massive sandstones with "dish" structure B2 Massive sandstones without "dish" structure.
BEDS CAN REASONABLY BE DESCRIBED USING THE BOUMA SEQUENCE	FACIES C -- Medium to fine sandstones -- classical proximal turbidites beginning with Bouma's division A. FACIES D -- Fine and very fine sandstones, siltstones -- classical distal turbidites beginning with Bouma's division B or C. C-D FACIES SPECTRUM -- can be described using the ABC index of Walker (1967). FACIES E -- Similar to D, but higher sand/shale ratios and thinner more irregular beds.
BOUMA SEQUENCE NOT APPLICABLE	FACIES F -- Chaotic deposits formed by downslope mass movements, e.g. slumps. FACIES G -- Pelagic and hemipelagic shales and marls -- deposits of very dilute suspensions.

and C should be found in the middle fan, depositional lobe area (suprafan) on a submarine fan (Figure 29). The suprafan area is characterized by depositional lobes surrounding distributary channels. The channels will contain coarser sediments than the lobes. As a layer of sediment is built up in the channel and in the lobe adjacent to it and further deposition is restricted, there is a tendency for the channel to migrate laterally, seeking a lower elevation. Thus, any given location (e.g. location X in Figure 29) may be characterized by coarse sands at time A, if it coincides with a distributary channel. Later, at time B, finer grained, overbank sediments may dominate because the channel has migrated out of the area. Still later, the channel may migrate back and coarse sediments return. The channel deposits are generally both thicker and coarser-grained than those of the distributary lobes.

In a prograding fan situation, the sediments at location X (Fig. 29) will exhibit a general thickening upward sequence. The reason is that, whereas location X once lay on the very fringes of the fan and received thin deposits of overbank and lobe sediment periodically (when the fan was small), it now is covered by middle fan deposits, some of which are thick channel sands. Sometime in the future it will be covered by the very thick deposits of the inner fan area.

It is suggested that the thickening upward sequence in the sand beds described in the Tyler Formation (STRATIGRAPHY AND LITHOLOGY) is a result of the processes cited above. Even though most of the beds exhibit constant lateral thickness in outcrop (though exposure is small)

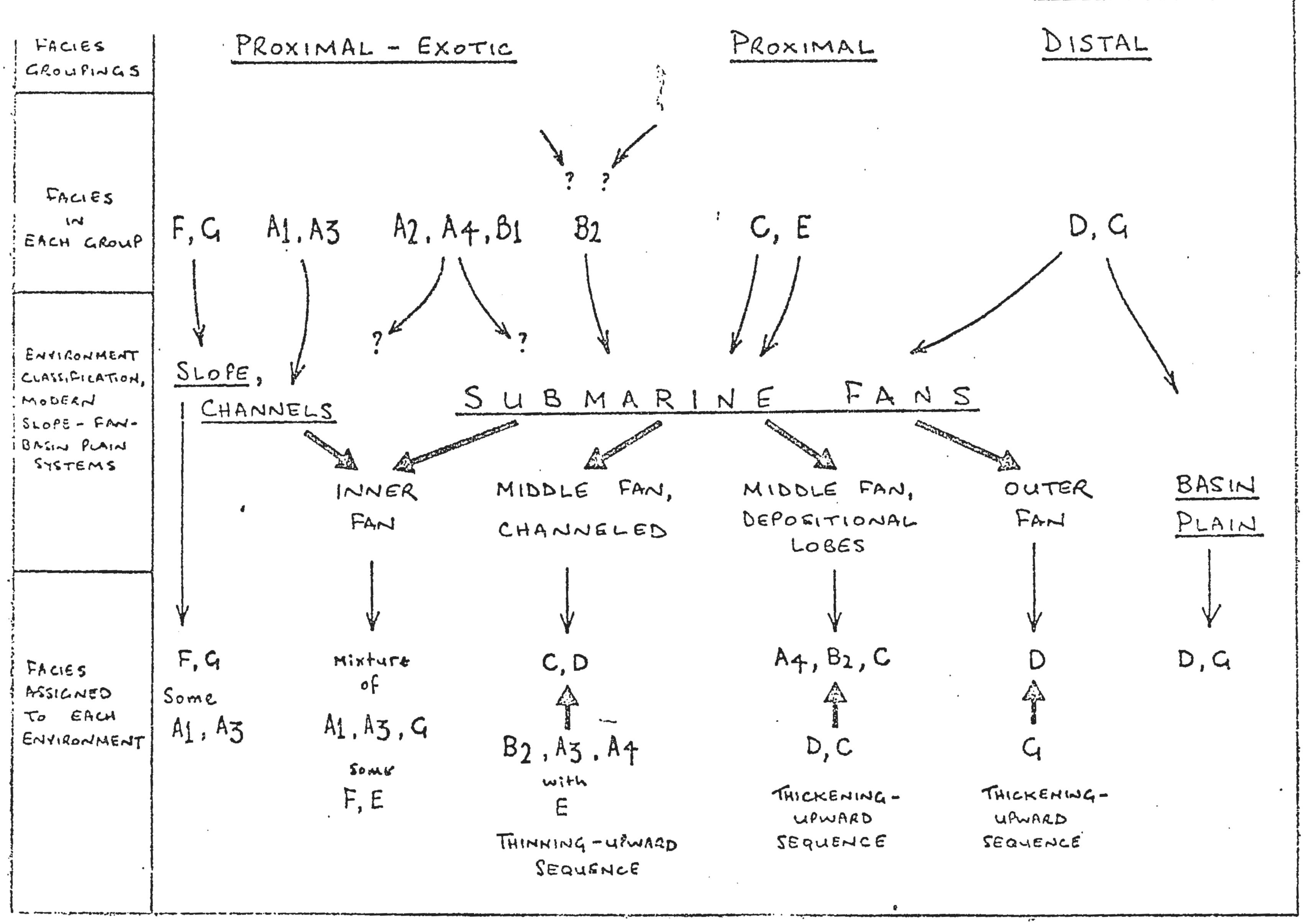


Figure 28--FACIES ASSOCIATIONS-TURBIDITES AND DEEP-WATER SEDIMENTATION
(from Middleton and Bouma, 1973).

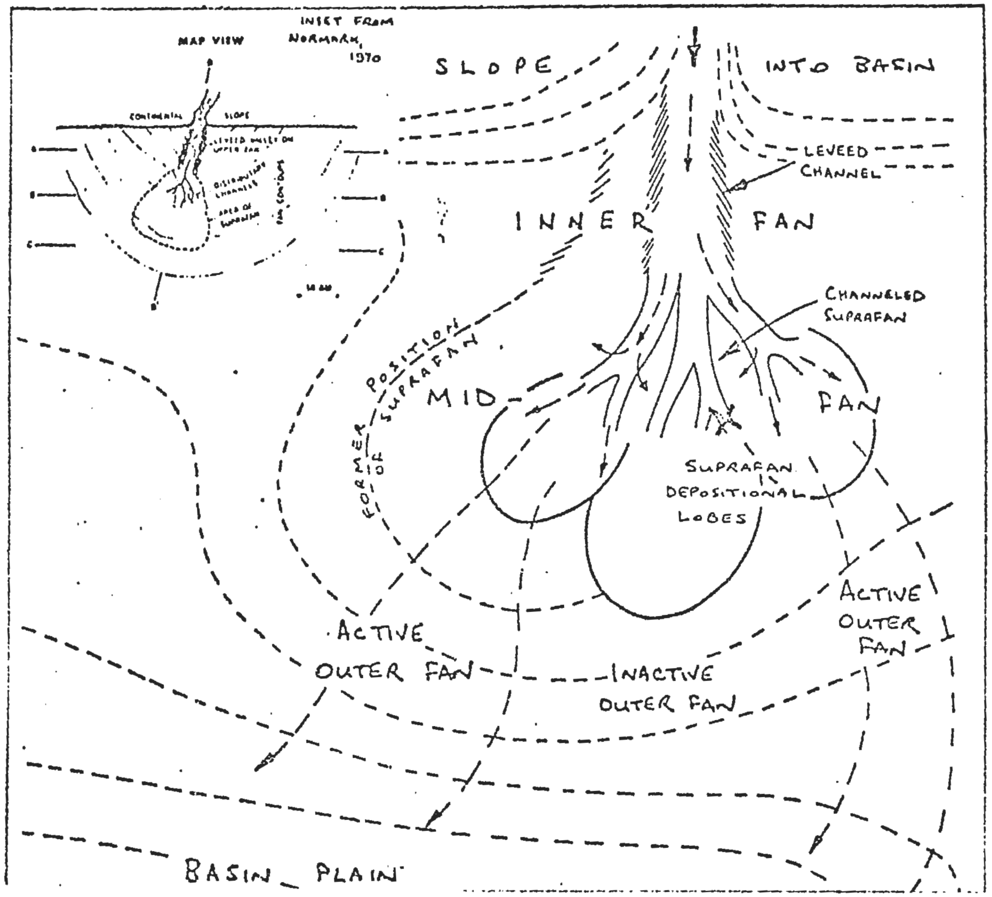


Figure 29--IDEALIZED SLOPE-FAN-BASIN FLOOR SYSTEM, SHOWING RELATIONSHIPS BETWEEN THE VARIOUS FACIES ASSOCIATIONS, AND POSSIBLE PALEOCURRENT DIRECTIONS. Zonation of the fan into Upper (inner), Middle and Lower (outer) segments is shown, with a suprafan developed in the middle fan area. The suprafan itself is subdivided into an upper channelized segment and a lower segment consisting of depositional lobes. (from Middleton, Bouma et al., 1973).

there is a hint of lensing which would characterize channel deposits (Figure 30). Because of the similarity between the Tyler and the theoretical facies and facies sequences of a submarine fan, it is suggested that that part of the Tyler Formation which was studied in detail (the measured section area) was deposited as part of a submarine fan complex.

Figure 30--PHOTO SHOWING LENSING OF A THICK GRAYWACKE BED (ARROW) SUGGESTING A POSSIBLE CHANNEL-FILL ORIGIN FOR THE BED.
(SE 1/4, NE 1/4, S14, T46N, R2E, Hurley, Wisconsin)

Paleocurrent Indicators

Paleocurrent indicators in the Tyler Formation are grouped into two categories, interstratal (between beds) and intrastratal (within beds). Included in the first category are groove casts, flute casts, "negative grooves" (ridge molds), flame structures, miscellaneous sole marks and ripple marks. Intrastratal indicators include cross-bedding and parting lineation.

One hundred and nine beds contained paleocurrent indicators. Some of the indicators are of questionable authenticity but general paucity of structures made each observed occurrence important to the final summary. Twenty-six of the outcrops yielded all of the paleocurrent indicators measured in this study.

Interstratal Paleocurrent Indicators

Interstratal paleocurrent indicators have been reported in the Tyler Formation (Felmlee, 1970; Komatar, 1972; Cooper, 1973) but have not been described in detail, nor have their orientations been systematically recorded. The most abundant of these structures are sole marks, with flute casts and groove casts the most useful for paleocurrent reconstruction. Sole marks are preserved as features of positive or negative relief on the bottoms of sand units which overlie argillaceous beds. Sand-mud interbedding is abundant in the Tyler Formation, but sole marks are not. Part of the reason may be the possible grain-flow origin of many of the sand beds. Although sole marks occur on the bottoms of grain-flow deposits, their irregularity (Stauffer, 1967) precludes the measurement of paleocurrent direction. In this study, sole marks that yield paleocurrent data are thought to be indicative of turbidity current deposition. This does not seem unreasonable since the great majority of the sole marks observed were recognizable groove casts or flute casts, known to characterize the soles of turbidites.

A correlation between thickness of sole marked sand beds and type of sole mark was noted (see Fig. 31). Flute casts generally occur on the soles of thickly-bedded units whereas groove casts occur on thinner units. This has been previously noted by Walker (1967). The average thickness of a flute-casted bed is 50 cm. with a range from 17.5 to 97 cm. (only one bed was less than 30 cm.). The average thickness of a groove-casted bed is 19 cm. with a range from 5.5 to 60 cm. (only one groove-casted bed was thicker than 25 cm.).

Groove casts are the result of dragging of a tool along the surface of a mud bed by a passing current. The tool scribes a groove parallel to current flow in the surface of the bed. Rapid burial, probably by the sediments of the same current which scribed the groove, preserves the structure as a feature of positive relief on the sole of the overlying bed (Fig. 32). The most common tools (especially in Precambrian rocks) are probably lutite clasts (Middleton and Bouma, 1973; Dzulynski and Walton, 1965). Such clasts have been found at the ends of some grooves in other formations (Dzulynski and Radomski, 1955; Glaessner, 1958) but not in the Tyler.

Twenty-four beds exhibited groove casts on which current directions were measured. Measurement of groove casts yields two possible flow directions, toward an azimuth parallel to the groove or toward its

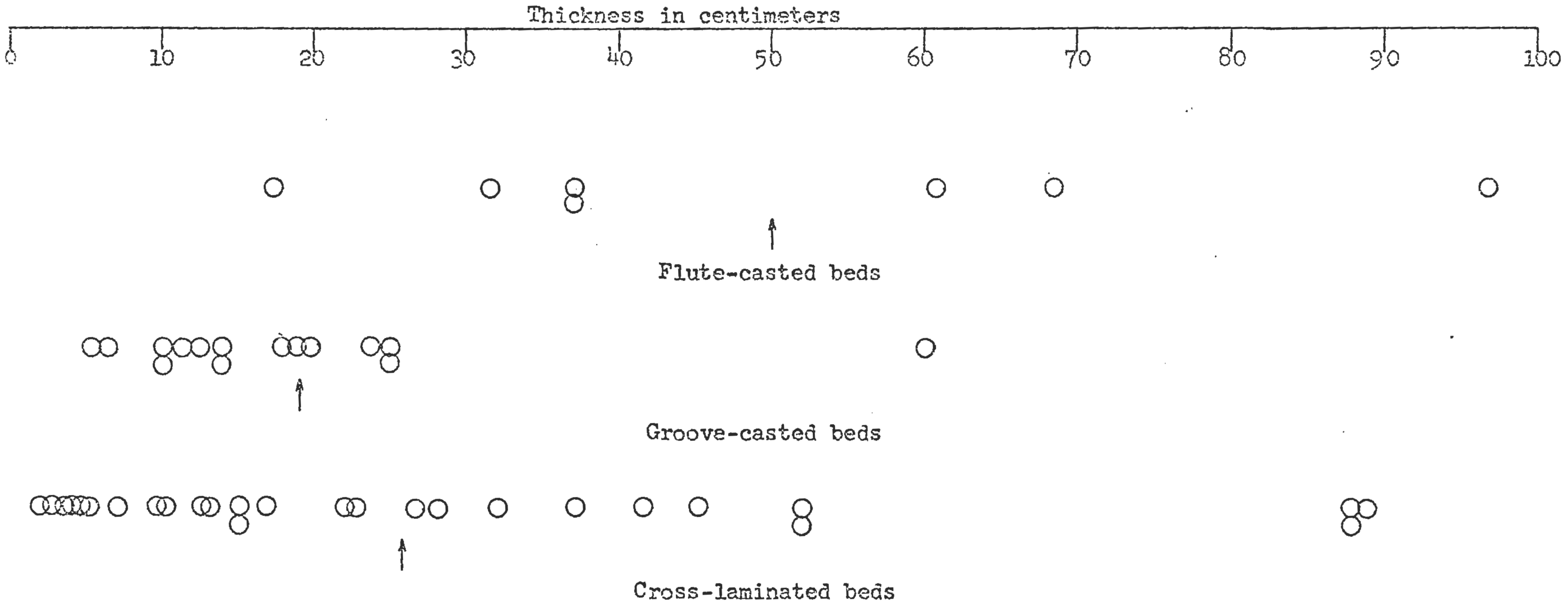


Figure 31--CORRELATION BETWEEN BED THICKNESS AND TYPE OF PRIMARY SEDIMENTARY STRUCTURE. Note that all flute casts found in this study (with one exception) occured on the soles of beds greater than 30 centimeters thick, while all groove casts (with one exception) were found on beds less than 26 centimeters thick. Cross-bedding appears to be independent of bed thickness. Arrows show average thicknesses of fluted, grooved & cross-laminated beds.

Figure 32--NUMEROUS GROOVE CASTS ON THE SOLE OF A GRAYWACKE BED. Scale in centimenters. (NE corner, SE 1/4, S31, T46N, R2E)

back azimuth. Such measurements are termed bipolar since the current could have been flowing in either of two directions. (Unipolar paleocurrent indicators, like cross-bedding and flute casts, yield a single azimuth.) One measurement was taken per sole. Groove casts were, in general, few in number on any single sole and widely spaced. Table 11 summarizes the dimensional characteristics of those groove casts which were measured.

Flute casts (Fig. 33) are the result of filling of scour marks by sand. The scour marks in the underlying mud unit were caused by turbulent eddies in the current which deposited the overlying sand unit (Allen, 1971a). Ten occurrences of flute marks were recorded and

TABLE 11--SOME MEASURED DIMENSIONS OF SOLE MARKS

	Length (cm)	Width (cm)	Relief (cm)
Groove casts	5	3	1
	6	.5	.1
	5	1	.5
	10	2	1
	14	3.5	.7
	9	5	.7
	3	1	.5
	12	.3	.1
	?	2	1
	?	5	1.5
	?	11	2
	?	9	.7
Flute casts	13	5	1.8
	?	6	1
	15	10	1.5
	12.5	8	1
	?	4	1
	25	15	2

Figure 33--LOADED FLUTE CAST ON THE SOLE OF A GRAYWACKE BED. Note that the beak of the flute (nearest the eraser end of the pencil) is turned under, suggesting soft sediment deformation of the original scour mark.
(SE 1/4, NE 1/4, S14, T46N, R2E near Hurley, Wisconsin)

current directions were inferred from their morphologies. Flute casts yield unipolar measurements. Dimensional characteristics of a few measurable flute casts are recorded in Table 11.

"Negative groove" is a purely descriptive term used to describe several elongate sole marks which do not resemble other marks with which the author is familiar. These marks are long, parallel striations on the soles of sand beds (Figures 34 and 35) which are characterized by negative relief rather than positive relief as are groove casts (Figure 32). They appear to be molds of what were once ridges on the tops of the underlying mud beds rather than casts of grooved surfaces. Figure 36 shows profile views of groove casts (A), "negative grooves" (B), and dendritic structures (C) which bear a strong resemblance to the negative grooves in profile (see Kuenen, 1957; Dzulynski and Walton, 1965). Dzulynski and Walton (1965, Figure 146) produced, by experimental means, some structures very similar to negative grooves.

Relief of the negative grooves is 0. 5 cm. or less and width is less than 1 cm. Longitudinally, they usually continue for the length of the bedding plane exposure although some are truncated (e. g. Figure 34). It is this irregular truncation which precludes the possibility that these features are several closely spaced groove casts. The intergroove spacing is variable from less than 1 cm. to 5 cm. Sometimes two negative grooves converge.

Negative grooves yield bipolar paleocurrent readings. It has been suggested by Dzulynski and Walton (1965) that the convergence of two

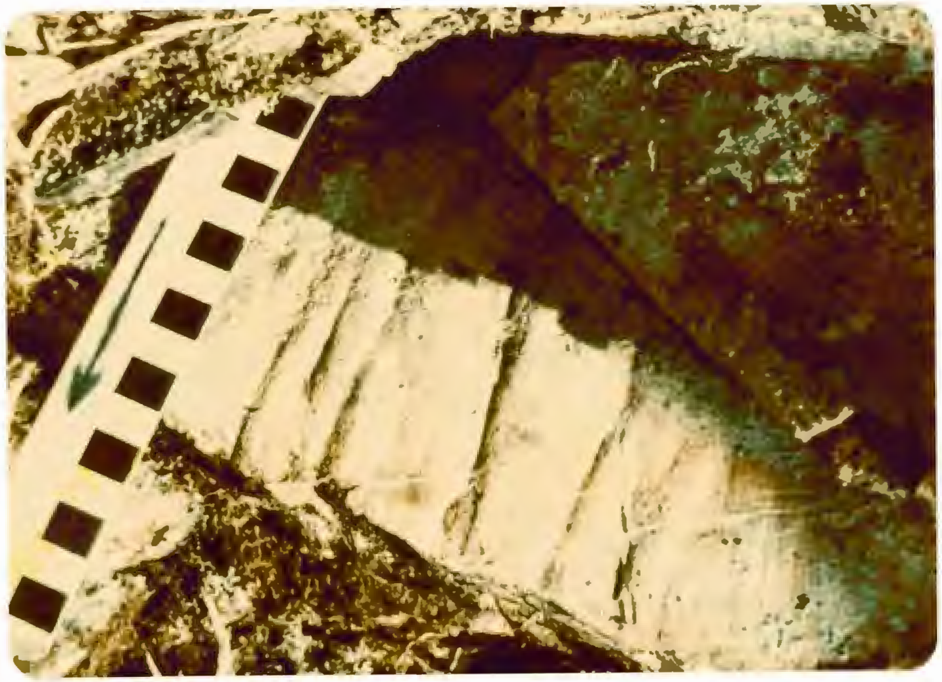


Figure 34--NEGATIVE GROOVES (RIDGE MOLDS) ON THE SOLE OF A GRAYWACKE BED. Note discontinuous nature of the two grooves nearest the scale and also the irregular spacing of the grooves. Scale in centimeters. (NW 1/4, NE 1/4, S19, T45N, R1E, Upson, Wisconsin)

Figure 35--NEGATIVE GROOVES (RIDGE MOLDS). Note the irregular spacing of the grooves (suggesting these are not ripple casts) and the bifurcation of the groove at the right end of the scale. Scale in centimeters. (NE corner, SE 1/4, S31, T46N, R2E)

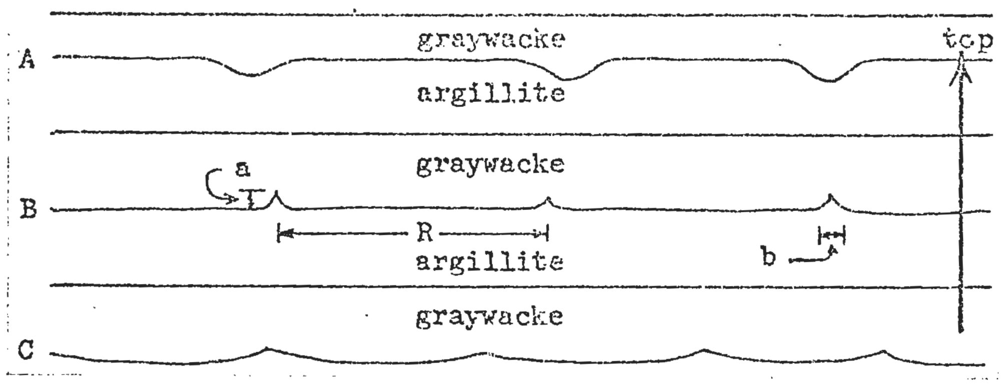


Figure 36--SCHEMATIC DIAGRAM OF PROFILES OF GROOVE CASTS (A), NEGATIVE GROOVES (B), AND DENDRITIC STRUCTURES (C). a = negative groove relief, b = negative groove width, R = intergroove spacing.

ridges is generally in a down current direction, but this is not always the case.

It seems likely that some of the negative grooves may be tectonic in origin. The very closely spaced species of ridges and furrows pictured in Figure 37 are especially thought to be pseudo-sedimentary, tectonic structures. At one outcrop in particular, some of the closely spaced ridges showed a parallel relationship to slickensides. A few of the widely spaced ridges were parallel to bedding-cleavage intersections (see Figure 38). Others bear no relationship to structural features whatsoever and these are considered primary.

Other sole marks noted include one prod mark, several saltation marks (skip casts, see Figure 39), chevron flutes (Figure 40), longitudinal ridge and furrow (Figure 41) and many irregular loaded surfaces (Figure 42). The ridge and furrow, skip casts and prod mark yield paleocurrent orientation, the latter two giving unipolar measurements.

Figure 37--SMALL NEGATIVE GROOVES (RIDGE MOLDS) WHICH WERE FOUND TO BE PARALLEL TO SLICKENSIDES UPON CLOSE EXAMINATION. (NW 1/4, NE 1/4, S19, T45N, R1E, Upson, Wisconsin)

Figure 38--PSEUDO-SEDIMENTARY, TECTONIC STRUCTURES WHICH RESEMBLE NEGATIVE GROOVES (RIDGE MOLDS) ON THE SOLE OF A GRAYWACKE BED. Note the parallelism of the "sole marks" and slaty cleavage (parallel to scale) in the bed immediately below. Scale in centimeters.
(same location as Figure 37)

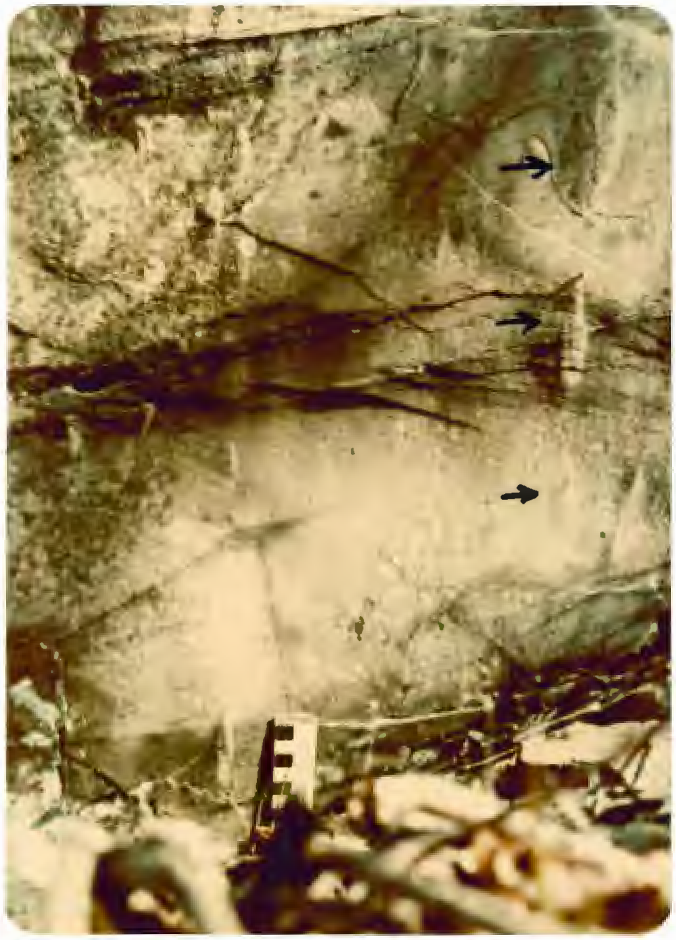


Figure 39--SKIP CASTS OR SALTATION MARKS (ARROWS) ON THE SOLE OF A TWO METER GRAYWACKE. Current movement from bottom to top as inferred from increasing size of casts toward top of photo. Each time the saltating tool hit the underlying mud bed it lost energy to the bed, making increasingly deeper and wider depressions in the mud as it bounced along (from bottom to top). Scale in centimeters. (same location as Figure 37)

The longitudinal ridge and furrow may give unipolar measurements as well since, like dendritic ridges, they tend to coalesce down current (Dzulynski and Walton, 1965).

Ripple marks, described in detail in the preceeding Description of Internal Bedding Features, were generally not useful for paleocurrent

Figure 40--MISCELLANEOUS SOLE MARKS. Note "chevron flutes" at far right. Current movement from top to bottom. Scale in centimeters. (same location as Figure 37)

reconstruction. Measurements of current orientation on the poorly preserved linguoid ripples were not good. The straight-crested ripples yielded poor measurements.

Summary diagrams of interstratal paleocurrent indicators are presented in Figure 43. Statistical and graphical tests of significance suggest that only flute casts, groove casts and negative grooves yield statistically valid vector means. A discussion of vector analysis technique and tests of significance will be found in the section entitled Paleocurrent Analysis.

Discussion of Interstratal Paleocurrent Indicators

Interstratal paleocurrent indicators are not abundant in the Tyler Formation even though cyclic sand-mud interbedding is ubiquitous. The

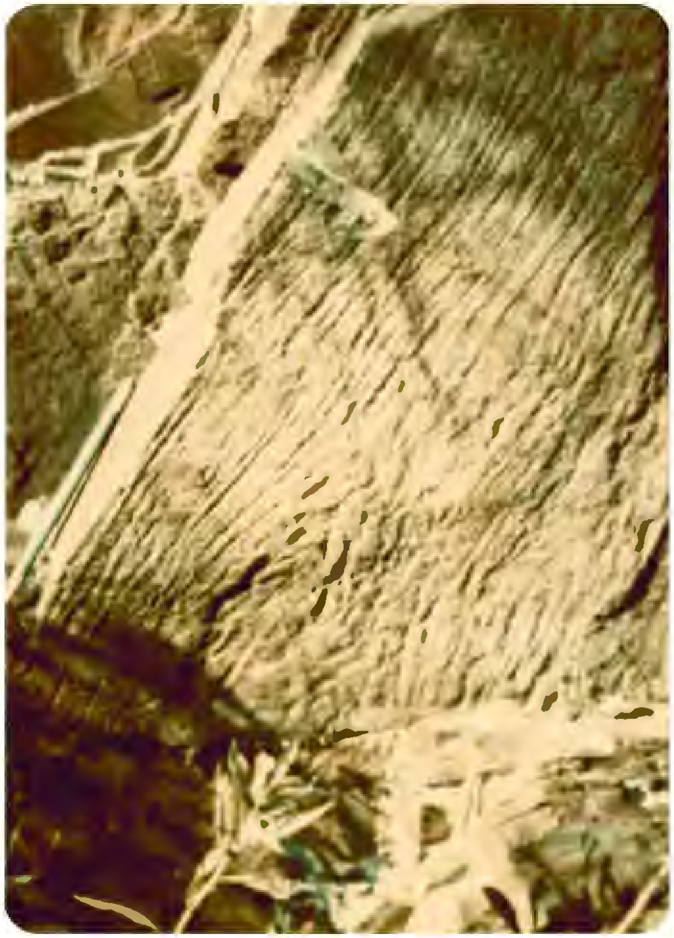
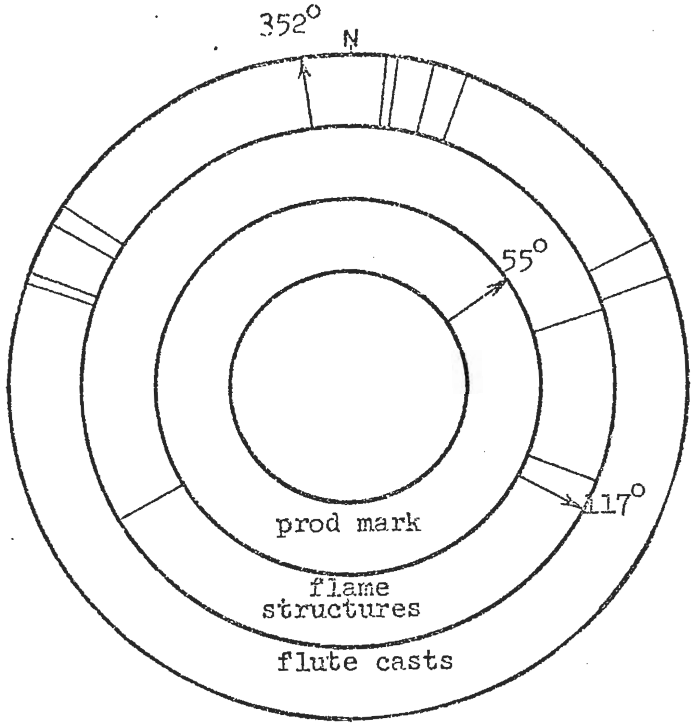
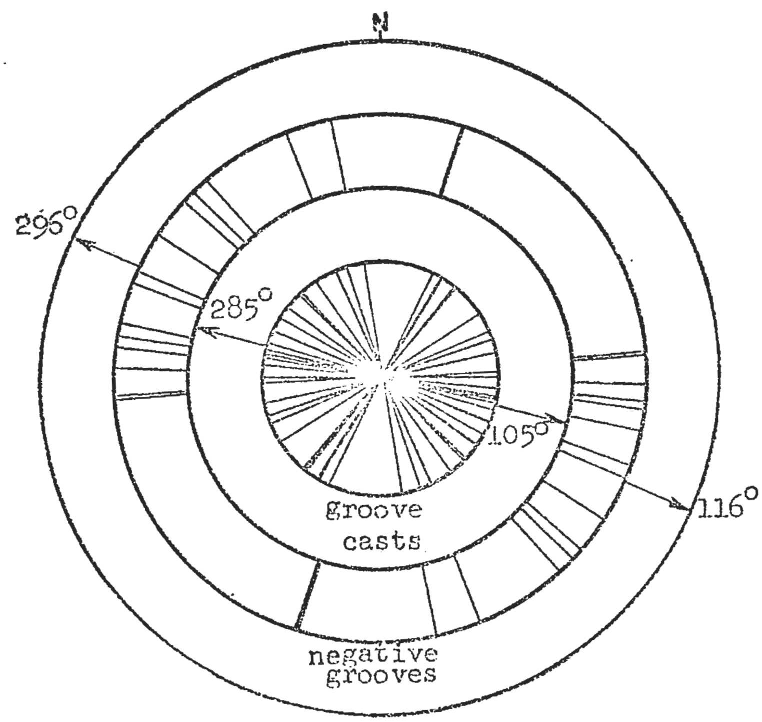


Figure 41--LONGITUDINAL RIDGE AND FURROW STRUCTURE (terminology after Dzulynski and Walton, 1965). Current movement parallel to pencil. (SE 1/4, NW 1/4, S28, T46N, R2E)

Figure 42--MISCELLANEOUS SOLE MARK. Irregularly loaded surface. Scale in centimeters.



UNIPOLAR INDICATORS



BIPOLAR INDICATORS

Figure 43--SUMMARY DIAGRAMS OF INTERSTRATAL PALEOCURRENT INDICATORS (arrows show vector means)

reason for this anomaly probably lies in the depositional mechanisms of of the sand units. More than 58% of the sand and silt beds could be of grain-flow origin and paleocurrent indicators are not common in grain-flow deposits.

The apparent relationship between bed thickness and type of sole mark (Figure 31) is probably not a relationship between those two parameters so much as a correlation between each of them and a third parameter, proximality to sediment source. Many authors (especially Walker, 1967) have pointed out an inverse relationship between distance from source and bedding thickness (closest beds being thickest). Also noted is an increase in tool mark abundance and a decrease in scour mark abundance away from the sediment source. It is expected that proximal currents (currents very close to their initiation site) are more competent than distal currents and can therefore deposit greater amounts of sediment after scouring underlying muds. Thus, thickness and presence of sole marks are not so much a function of each other as they are a function of current competency which is partially a function of proximality and partially a function of basin bottom (or submarine fan) morphology.

Whether or not an unlithified lutite clast (see Description of Internal Bedding Features) is sufficiently competent to scribe grooves in other soft argillaceous sediments is probably open to question. The writer is not aware of studies concerned with the "scribing tools" aspect of tool marks although there are several dealing with the "how to" of these sole marks. Presumably, a soft lutite clast, if moving fast enough, can scrape grooves in other soft sediment. This is especially true when the tool is a cohesive mud clast. The dimensions of groove casts in the Tyler Formation certainly do not rule out the possibility that clasts of similar size and shape as those presently found preserved in the sandstone beds could have been the scribing tools.

The origin of negative grooves (ridge molds) is problematical. They could be primary sedimentary structures, secondary soft sediment deformation structures, tectonic or even biologic. However, they are too regular in appearance to be attributed to load casting (soft sediment deformation) and perusal of the several plates which accompany the text in Crimes and Harper (1970) and Hantzschal (1975) failed to reveal any ichnofossils morphologically similar to negative grooves. It is believed that some of the closely spaced grooves are tectonic (see above) and the more widely spaced, bifurcating grooves similar to those described by Dzulynski and Walton (1965) are primary. Perhaps they are the molds of ridges in the interlobe (cleft or tunnel) areas at the head of a lobate turbidity current (see Figure 44 and Allen, 1971b). Such ridges could be constructed by ambient waters forced away from encroaching lobes. These secondary spiraling currents would be asymptotic to Allen's tunnel center streamlines (Figure 44) as they approach tunnel center from both sides. The result of their convergence would be a mud ridge parallel to paleocurrent direction and somewhat analagous (genetically speaking) to the larger scale sand ridges described earlier.

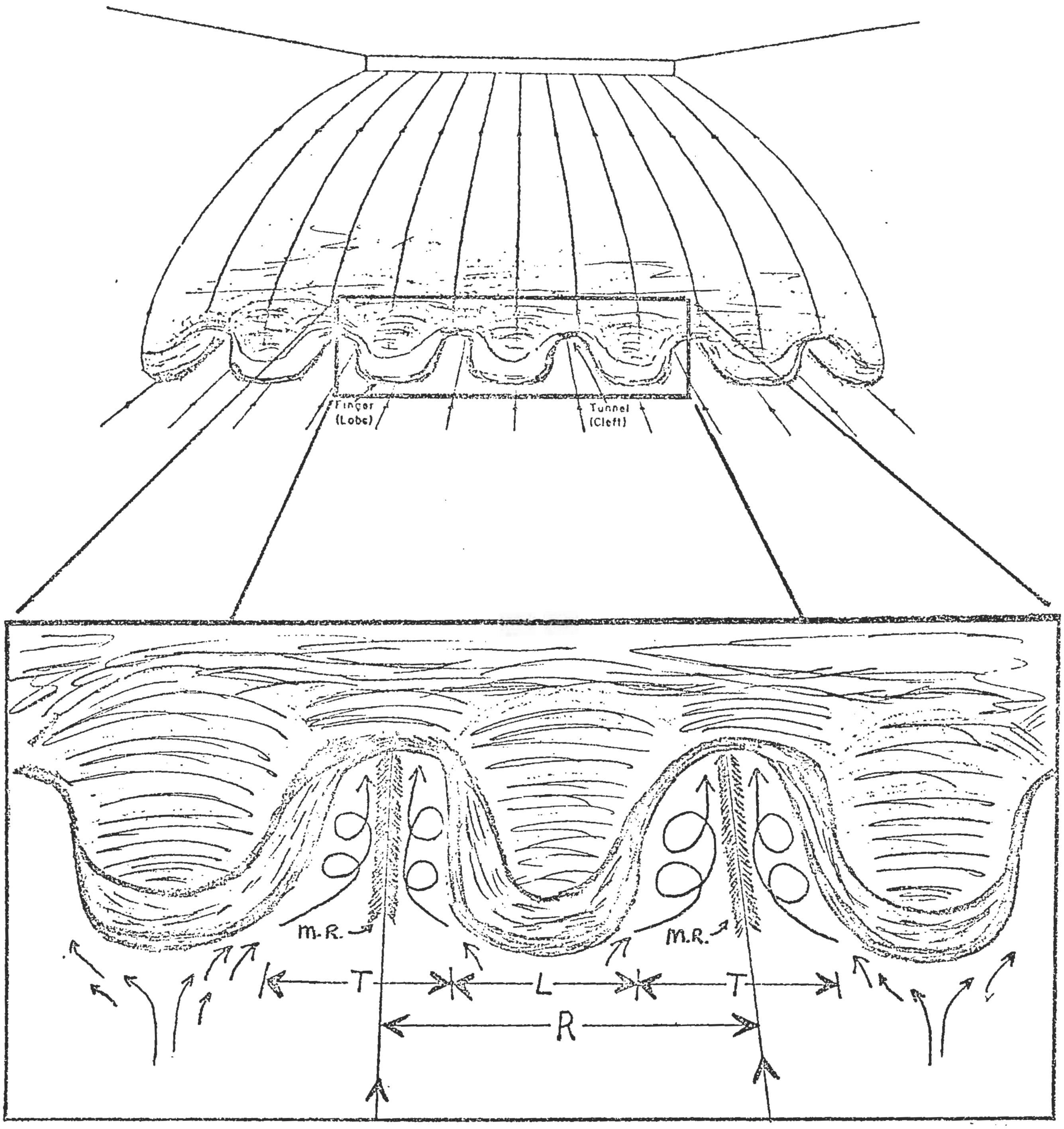


Figure 44--FRONT VIEW OF TURBIDITY CURRENT HEAD SHOWING TUNNELS AND FINGERS. (after Allen, 1971b) Helicoidal currents entering tunnels pile mud into ridges (M. R.). L=lobe width, T=interlobe width and R=interridge spacing. Note that arrows show relative movement only. The currents entering the tunnels are actually moving in the same direction as the turbidity current, but more slowly.

Such ridges may be preserved only if undisturbed by turbulence within the current and if quick burial by sediment occurs. This suggests that either the lobes of the head did not migrate across the tunnel-center ridges or that the current was very weak, with little internal turbulence. But, Allen (1971b) suggests that lobes probably do migrate transversely as the current moves. The alternative explanation, that the current was lacking sufficient internal turbulence or competency to destroy the tunnel center ridges, seems to be substantiated by the absence of other sole marks (especially of scour marks) on the same bedding planes.

The spacing of negative grooves is of the same order of magnitude as interlobe spacings (between sets of sole marks) measured by other authors. Interlobe spacing varies from 1 m. (Kuenen, 1957), 30-40 cm. and 2-4 cm. (Dzulynski, 1963, in two separate examples) to 25 cm. (Dzulynski and Walton, 1965). Negative groove (ridge-mold) spacing is slightly smaller and varies from 6 cm. to less than 1 cm.

If negative grooves are the molds of tunnel-center ridges, then certain inferences can be made about the spacing of the tunnels in the current head as it passed through the area. Calculations based on the spacing of ridge molds yield current velocity of 21 cm. sec.$^{-1}$ and primary slope of 0^{o}9.8' (see Appendix D). The calculated slope compares favorably with observed bottom slopes on modern submarine fans (Table 12).

TABLE 12--DIMENSIONS OF REPRESENTATIVE SUBMARINE CANYON-FAN SYSTEMS OF THE WORLD (from Middleton and Bouma, 1973)

Name of canyon-fan system[1]	*Dimensions*[2]						*Gradient*		*Gradient*	
	canyon (length)		*fan (length)*		*fan (width)*		*canyon*		*fan*	
	nm	*km*	*nm*	*km*	*nm*	*km*				
1 Bering (U.S., Alaska)	220	(407)	—	—	—	—	0°27′	1:125	—	-
2 Zhemchug (U.S., Alaska)	126	(233)	—	—	—	—	0°48′	1:71	—	-
3 Ganges-Bengal (India)	100	(183)	1,390	(2,570)	590	(1,091)	0°27′	1:128	04′	1:89
4 Congo (Africa)	120	(222)	150	(278)	100	(185)	0°34′	1:101	15′	1:22
5 Monterey (U.S., Calif.)	60	(111)	165	(305)	120	(222)	1°31′	1:38	14′	1:25(
6 Mississippi (U.S., La.)	120	(222)	120	(222)	80	(148)	30′	1:117	29′	1:12(
7 Astoria (U.S., Ore.)	62	(115)	90	(166)	55	(102)	57′	1:60	18′	1:19(
Willapa (U.S., Wash.)	60	(111)	—		—		58′	1:56	—	—
8 Hudson (U.S., N.Y.)	50	(92)	80	(148)	80	(148)	1°15′	1:46	23′	1:147
9 Rhone (France)	15	(28)	90	(166)	90	(166)	3°11′	1:18	24′	1:144
10 La Jolla (U.S., Calif.)	7.3	(14)	16	(30)	12	(22)	2°17′	1:25	1°04′	1:54
11 Redondo (U.S., Calif.)	8	(15)	4	(7)	6	(11)	2°12′	1:26	1°11′	1:48

[1] Data after: 1 = SCHOLL et al. (1970); 2 = SCHOLL et al. (1970); 3 = HEEZEN and THARP (1964); BATES et al. (1959); UCHUPI (1967); 7 = ROYSE (1964); GRIGGS (1969); 8 = HEEZEN et al. (1959); SHEPARD SHEPARD and BUFFINGTON (1968); 11 = BATES et al. (1959); EMERY (1960).
[2] fms = fathoms; nm = nautical miles.

SHEPARD and DILL (1966); 4 = BATES et al. (1959); SHEPARD and DILL (1966); 5 WILDE (1965). 6 – and DILL (1966); 9 = MENARD et al. (1965); SHEPARD and DILL (1966); 10 = SHEPARD and DILL (1966).

Summary of Interstratal Paleocurrent Indicators

In summary, the kinds of interstratal paleocurrent indicators present in the Tyler Formation suggest deposition of the beds which contain them by a turbidity current mechanism. Vector means (Fig. 43) suggest currents moving to the WNW for tool marks and negative grooves and to the NNW for flute marks. Calculations based upon observed primary sedimentary structures suggest head heights of less than 10 cm. for at least some of the currents (Appendix D), velocities on the order of a few tens of centimeters per second (Appendices C & D) and a primary slope of 0^{O} 9. 8' (Appendix D).

Intrastratal Paleocurrent Indicators

This group of primary sedimentary structures includes cross-bedding and parting lineation. Cross-bedding has been previously reported by Cooper(1973) and Komatar (1972). A complete description of the cross-bedding found in the Tyler Formation is given in the section on Description of Internal Bedding Features. What remains here is to discuss the paleocurrent measurements made on the cross-sets.

Forty-five measurements were made on cross-sets. Their attitudes were plotted on a stereograph and rotated back to horizontal. No plunge (or gentle plunge) was assumed. A pole to the plane of the rotated cross-set in the direction of dip gave the azimuth of current movement, a unipolar measurement. Those azimuths are plotted in Figure 45. Statistical and graphical tests of significance (see

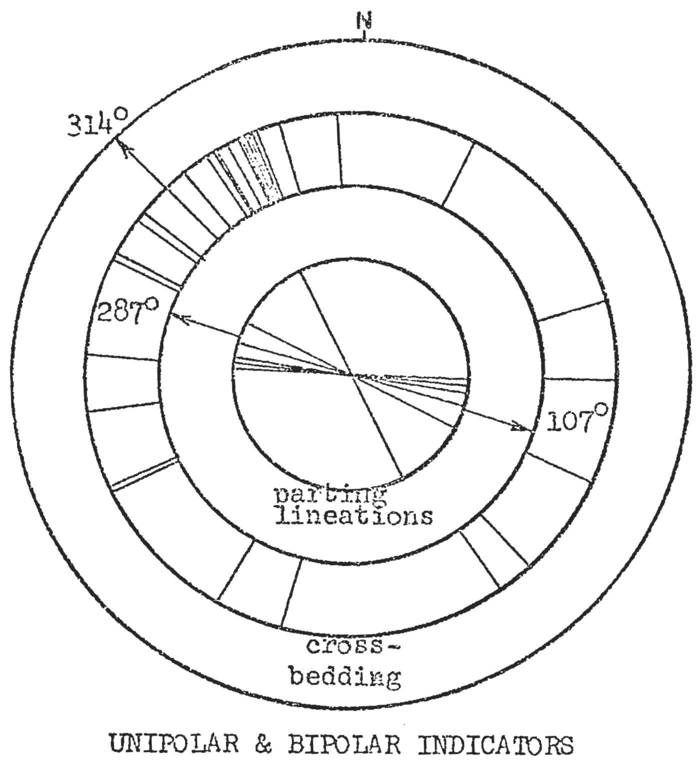


Figure 45--SUMMARY DIAGRAM OF INTRA-STRATAL PALEOCURRENT INDICATORS.
Parting lineations are plotted in inner circle.
Cross-beds are plotted in 3rd circle.
Arrows show vector means (bipolar for parting lineations and unipolar for cross-beds).

Paleocurrent Analysis) suggest that the vector mean for the cross-sets is reliable.

Parting lineation is the post lithification expression of the streaming of sand grains during deposition (McBride and Yeakel, 1963). Six bipolar measurements were taken on parting lineation in the Tyler Formation. Those measurements and their statistically valid vector mean are plotted in Figure 45.

Summary of Paleocurrent Indicators

Indicators of current movement found in the rocks of the Tyler Formation are of several types and include both interstratal and intra-stratal. All of the indicators are types commonly found in turbidity current deposits. Figures 43 and 45 summarize the orientations of those features measured in conjunction with this study. The currents which deposited the Tyler sediments moved from the east-southeast toward the west-northwest. A more detailed analysis follows.

Figure 46 is a summary of paleocurrent indicators measured in the Tyler and surrounding formations by several authors. There is general agreement between their measurements and those made in this study.

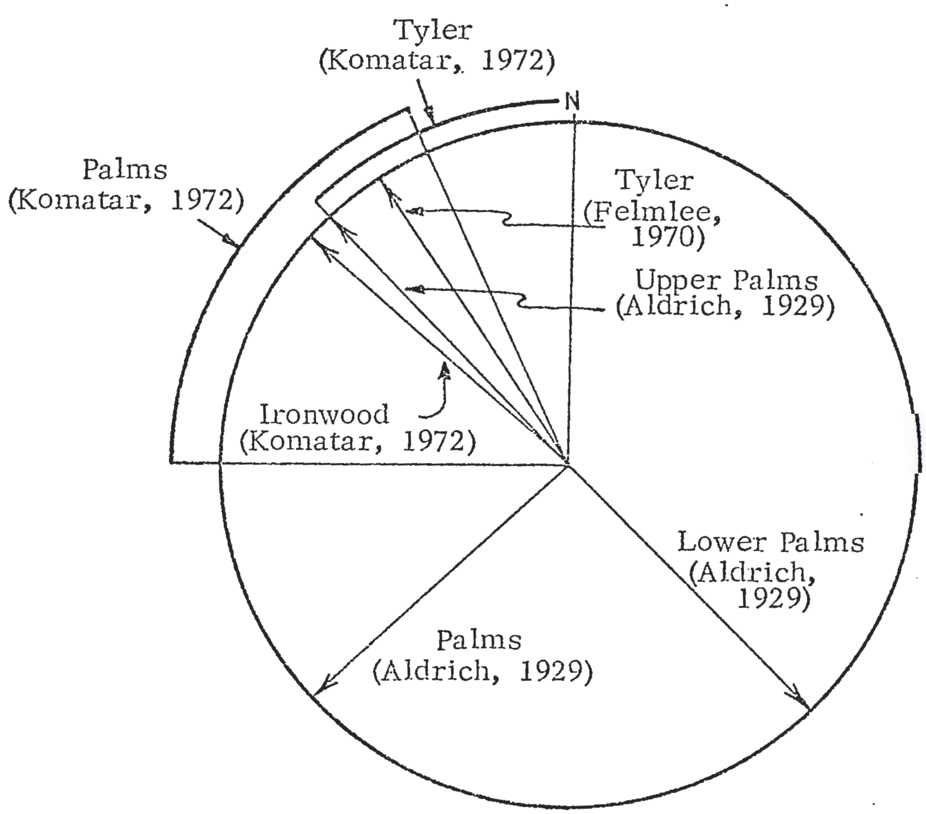


Figure 46--ORIENTATIONS OF PALEOCURRENT INDICATORS IN THE TYLER AND SURROUNDING FORMATIONS (from several authors, as shown)

Other Sedimentary (?) Structures

Other sedimentary structures noted in the Tyler Formation which are not easily defined as primary or secondary include convolute beds, load casts, ball and pillow structure and flame structures. Some others are not clearly sedimentary in nature. Small faults which offset a single bed only and cannot be traced into adjoining strata and which are thought to have formed before lithification of the sediments were found at the base of the formation. Concretions, leaching and cone-in-cone structures were noted elsewhere. Some concretions were centered around lutite clasts suggesting that the clast composition triggered precipitation of the concretionary mineral, which was calcite in the few cases examined. None of these features are particularly germane to the objectives of this thesis and were not studied closely.

Paleocurrent Analysis

Vector analysis of the paleocurrent data from the Tyler Formation (Appendix E) yields the vector means plotted in Figures 43 and 45. All vector means which are significant (Appendix E) are in close agreement and document movement toward the west-northwest. Azimuths of unipolar vector means include 314^{o} for cross-bedding and 352^{o} for flute marks. Azimuths of bipolar vector means include 285^{o} (-105^{o}) for groove casts, 296^{o} (-116^{o}) for negative grooves and 287^{o} (-107^{o}) for parting lineations.

The Tyler outcrop belt was subdivided into six geographic areas to look at lateral variation in paleocurrent trends (see Figure 47). Subdivision was based upon the amount of raw data per outcrop, an effort being made to include approximately the same number of readings in each area regardless of the number of outcrops thus included. Resultant vectors were then calculated for all unipolar data for each area and for all bipolar data for each area. The results are shown in Figure 47.

A plot of paleocurrent indicators vs. stratigraphic position was made to look for variation in current trend through time (Figure 48). Only paleocurrent indicators from the outcrops of the measured section were plotted since only at those outcrops can relative stratigraphic position be ascertained. Bipolar and unipolar vector means are shown, as are the arithmetic averages of those means. There seems to be no significant change in orientation of Tyler paleocurrents through time.

Discussion of Paleocurrent Analysis

Those vector means which are believed to be significant are all in close agreement with the single exception of the mean calculated from flute casts. This agreement documents movement of the currents which deposited the Tyler sediments from east-southeast to west-northwest. The flute casts (10 measurements) suggest movement toward the 352° azimuth (Figure 43), a full 60° away from the arithmetic mean (292°) of the other vectors. This discrepancy is considered too great to be explained by internal variation in flow direction of turbidity currents as

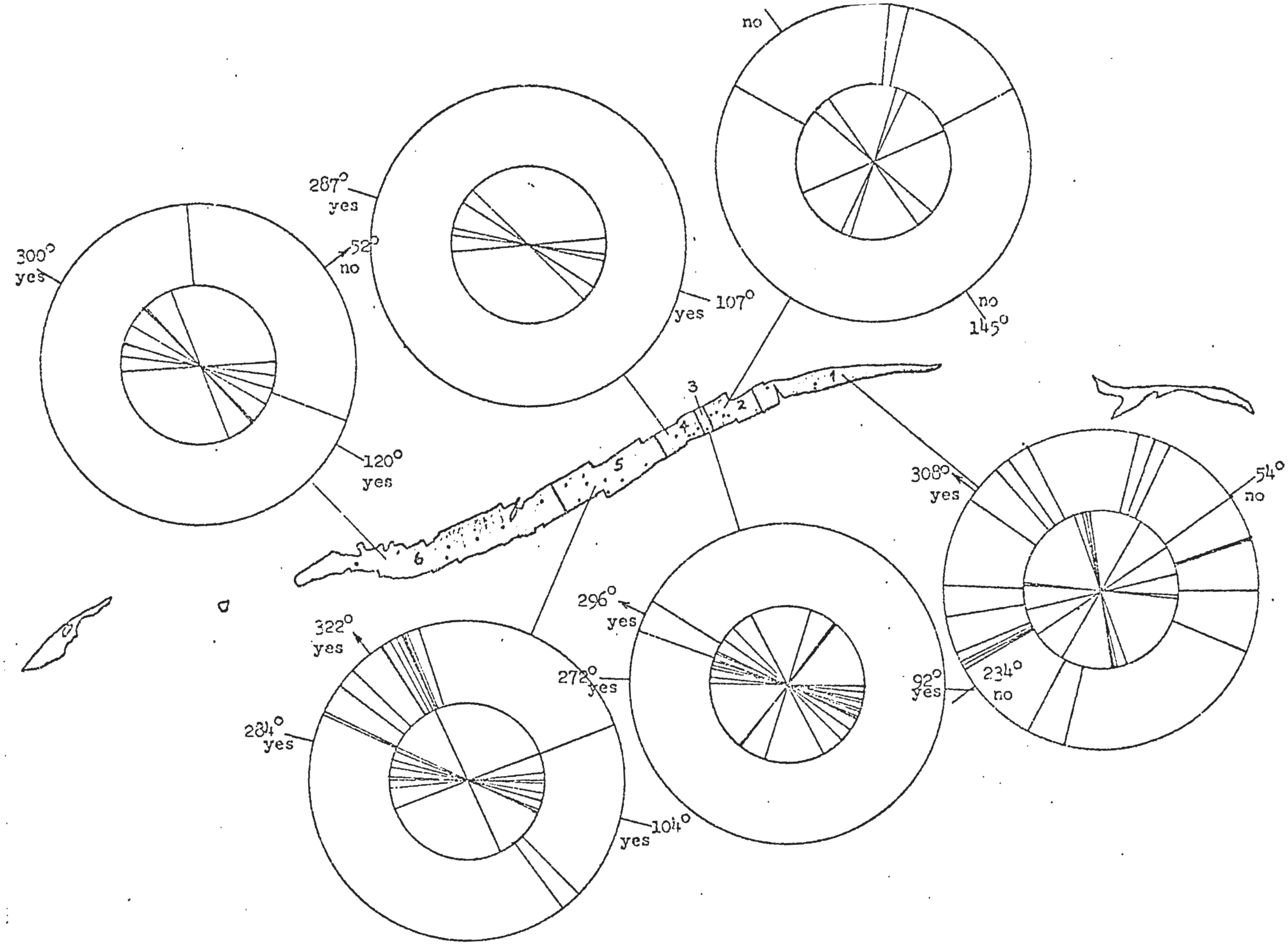


Figure 47--SUMMARY DIAGRAM OF PALEOCURRENT INDICATORS BY GEOGRAPHIC SUBDIVISION. Inner circles-bipolar data. Outer circles-unipolar data. Unipolar vector means shown with arrows. Bipolar vector means shown with bars. Yes or no indicates significance of vector mean according to F-test (see Appendix E)

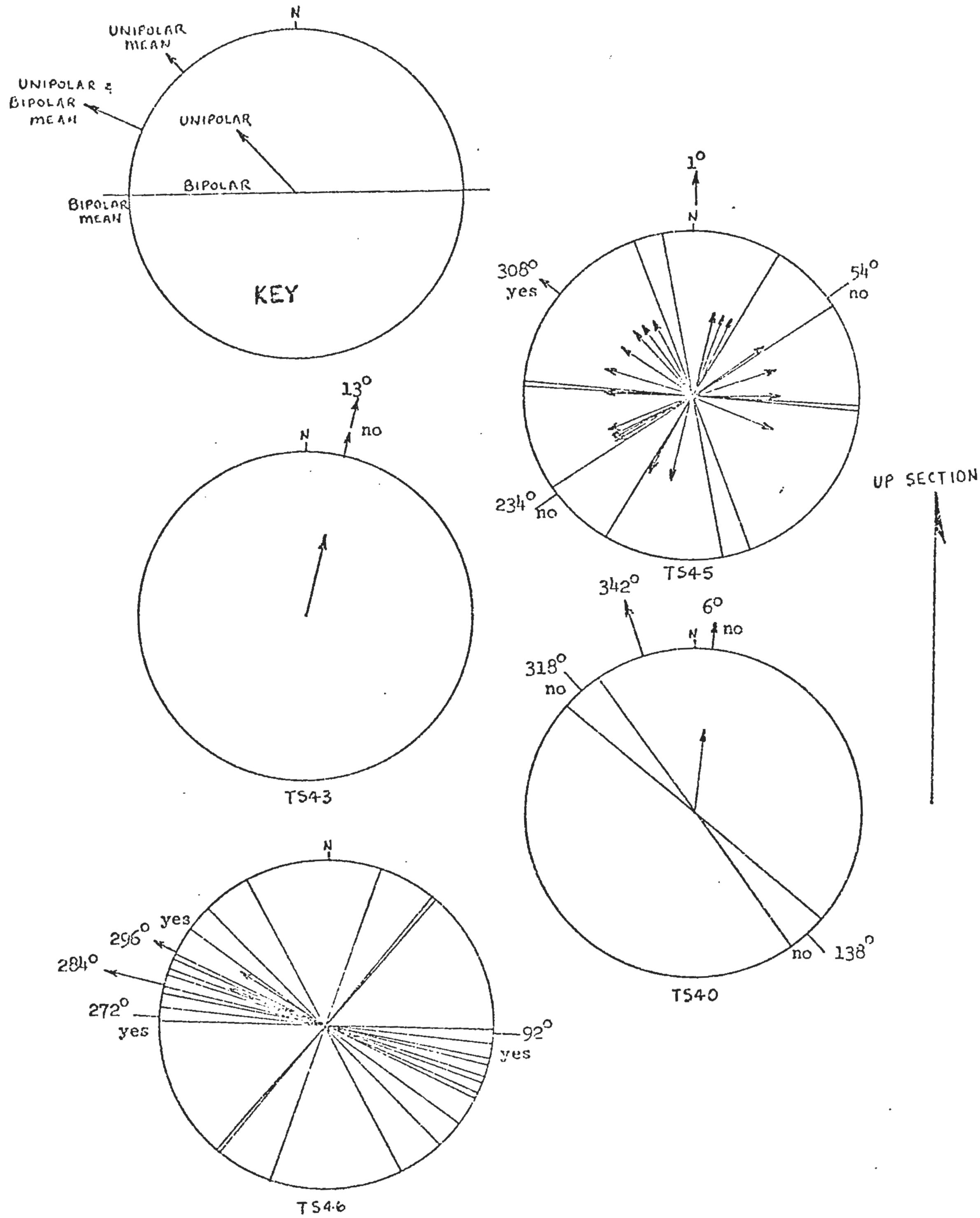


Figure 48--STRATIGRAPHIC VARIATION IN PALEOCURRENT TRENDS IN THE TYLER FORMATION. Arrows are unipolar data. Bars are bipolar data. Vector means are shown outside the circles (see key). TS numbers are outcrop numbers. Yes and no refer to F-test of significance. (see Appendix E).

they fan out over the depocenter. It is suggested that the discrepancy is the reflection of true variation in current trend due to geography and/or topography.

Figure 49 is a plot of interstratal paleocurrent indicators by geographic subdivision. Note the change in orientation of unipolar indicators (outer circle) between subdivision 3 and subdivision 1. The current direction rotates from NW (3) to N (2) to NE (1), the single SW plot in subdivision 1 being a flame structure the reliability of which is questionable (see significance tests - Appendix E). It is suggested that the prevailing currents were oriented differently in the east end of the study area than they were elsewhere or, alternatively, that there were two dominant current orientations. A single vector mean calculated for all the flute cast data would, of course, not show this feature. The vector mean can be likened in this instance to an arithmetic average which tends to obscure bimodal patterns.

A more reasonable analysis can be made by calculating three separate vector means for geographic divisions 1, 2, and 3. These vectors for the flute casts (Figure 49) document the shift in current direction toward the eastern end of the study area. Reference to field notes indicates that the flute cast at 288°, subdivision 1 is not a reliable reading. Eliminating it from the vector mean calculation for subdivision 1 would put the mean at azimuth 17° making the paleocurrent pattern shift even more clear than is shown in Figure 49. The 60° discrepancy suggested by Figures 43 and 45 is thus shown to be a mathematical

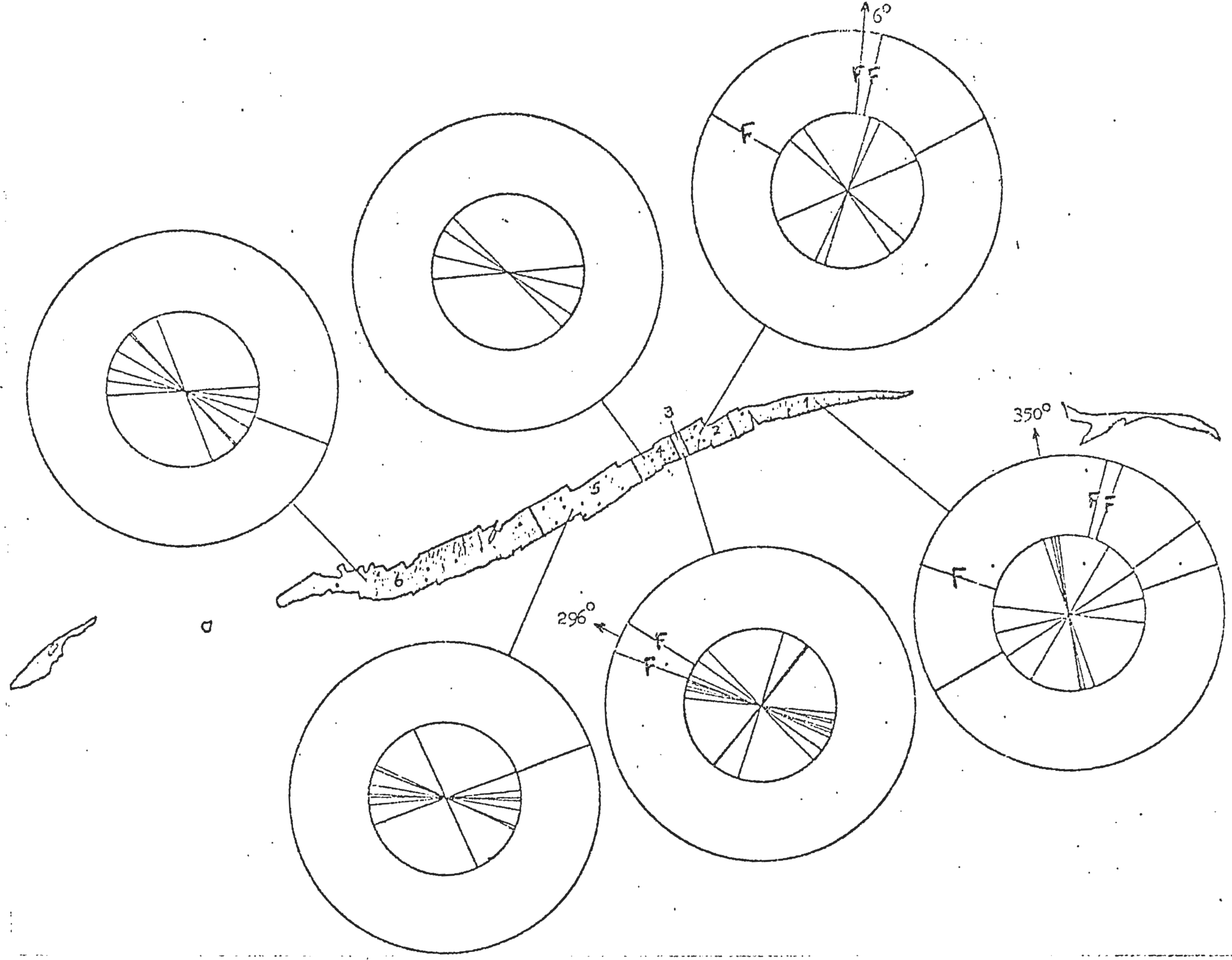


Figure 49--SUMMARY DIAGRAM OF INTERSTRATAL PALEOCURRENT INDICATORS BY GEOGRAPHIC SUBDIVISION. Bipolar data-inner circles. Unipolar data-outer circles. Flute casts marked F. Arrows are flute cast vector means.

discrepancy which probably reflects actual current trends.

The same shift in current should be noted in the bipolar indicators. Reference to the plots of individual readings (Figures 43 and 45) shows quite a few bipolar paleocurrent indicators in a NE - SW orientation. The vector means do not reflect this because the NW - SE indicators far out number the NE - SW indicators. Those calculated vector means are, therefore, not as sensitive to a few anomalous readings as is the flute cast vector mean where even one anomalous reading will significantly effect the calculation.

Figure 47 shows that all the paleocurrent indicators (both interstratal and intrastratal) reflect this easterly change (subdivisions 3, 2 and 1) from an otherwise very clear NW - SE orientation (subdivisions 4, 5, and 6). It is suggested that this current shift was due to local variation in basin bottom topography. The currents which deposited the rocks included in subdivisions 6, 5, 4 and 3 may have flowed down the northwest flank of the presumed submarine fan (see STRATIGRAPHY AND LITHOLOGY and Discussion of Bedding Features), while those in area 2 flowed north off the front of the fan and those in area 1 drained the northeast flank of the fan. This reasoning would tend to put the apex of the submarine fan somewhere in area 2.

Reconstruction of the local geography during the Late Middle Precambrian has suggested a highland to the south and east of the present Tyler outcrop belt. A basin margin slope would have been found at approximately the present position of the Early Precambrian-Middle

Precambrian contact. That fan became the locus of deposition of the Tyler sediments and probably lay at the base of that slope. Directions of current movement probably did not change significantly throughout Tyler time.

PETROLOGY

Ninety-four samples of the Tyler Formation rocks were selected for sectioning and microscopic study. The general guidelines for selecting those samples were the following objectives: to obtain a representative vertical and lateral sampling; to sample all lithologies; to sample beds with paleocurrent indicators so that composition could be tied to probable source areas. All of the thin sections were studied for compositional and textural character, but not all were point counted.

The major framework constituents are quartz, plagioclase and rock fragments. Grain size is highly variable, from silt to granules. Anything finer than .03 mm. is considered matrix; included here are fine chlorite, sericite, calcite, biotite, opaques, quartz and feldspar.

All of the rocks examined in thin section are graywackes (> 15% matrix) except for a few mudstones (> 75% matrix) according to the rock classification of Pettijohn, Potter and Siever, 1973. They are compositionally submature and texturally immature.

Framework Constituents

Quartz is the major constituent of the rocks of the Tyler Formation. It occurs as the most abundant of the detrital grain species, as secondary quartz veinlets and in at least one case as authigenic cement in an unusually clean (lacking muddy matrix) pocket of quartz sandstone. Framework quartz grains are both monocrystalline and polycrystalline and range in size from silt (< .0625 mm.) to granules (> 2 mm.). The smaller the grain size, the more dominant is monocrystallinity. Polycrystalline grains occur only rarely in Tyler siltstones.

The quartz grains are always very angular to subangular and range in shape from equidimensional to elongate. They show all variations of extinction from very sharp to extremely undulatory. The internal crystal boundaries of the polycrystalline types are sometimes smooth and show 120^{o} triple junctions but are most often irregular and in some cases sutured. Polycrystalline grains with irregular crystal boundaries which show elongation of individual crystals are classified as metamorphic rock fragments (see below).

Figure 50 shows a monocrystalline quartz grain with an optically continuous quartz overgrowth. It is hard to envision such a clean overgrowth as this occurring in a muddy rock. Also, other quartz grains in the same thin section do not show similar overgrowths. It is suggested, therefore, that this is a second cycle quartz grain which acquired the overgrowth while it was part of a cleaner quartz sedimentary rock. That rock was then subjected to erosion and this grain was redeposited as a framework grain of the graywacke in which it is now found. The very irregular edges of the quartz overgrowth are probably the result of chemical encroachment by the surrounding matrix.

The varieties of feldspar which are recognized in thin section include plagioclase, perthite, microcline and orthoclase. Most of the feldspar fragments are angular but some are subrounded. Alteration to sericite has tended to obscure crystal outlines. Often a feldspar grain

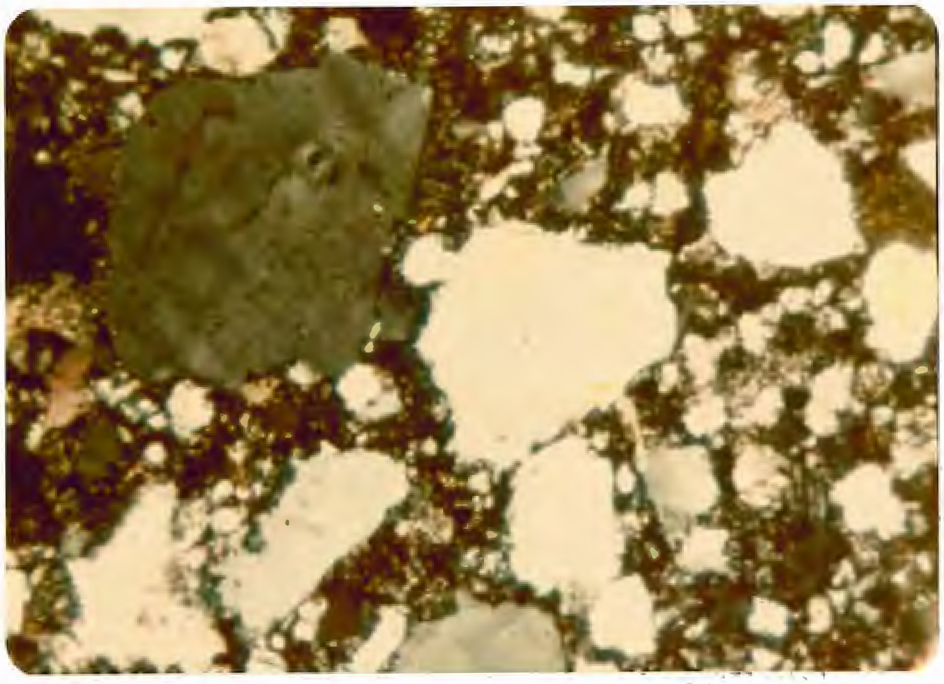


Figure 50--PHOTOMICROGRAPH OF GRAYWACKE WITH RE-WORKED MONOCRYSTALLINE QUARTZ GRAIN (CENTER). Note preserved outline of original grain (arrow) now surrounded by optically continuous overgrowth. Field of view approximately 2 mm. Crossed nicols.

is completely rimmed by sericite. Recognition of the feldspars where they have been largely altered to chlorite and sericite is difficult, but usually enough of the crystal remains so that sharp extinction of the entire grain permits distinction between felspars and aphanitic rock fragments, most of which also are altered. In other cases the feldspars are altered preferentially along cleavage traces so oriented sericite preserves the outline of cleavage fragments. Relatively fresh feldspars are also present.

Plagioclase is by far the most abundant feldspar. Both twinned and untwinned varieties are present, the former being the most common. Untwinned plagioclase is indistinguishable from orthoclase in thin section and the percentages of these minerals is based upon visual estimates made on thin section heels which have been stained for K-feldspar with cobaltinitrite. For point counting purposes all untwinned feldspar (with the exception of perthites) were counted in the orthoclase category. The resultant percentages of orthoclase feldspars were then adjusted downward to agree with the visual estimates, the excess being added to the plagioclase percentage. Normally zoned plagioclase is present in trace amounts (Figure 51). Some twinned plagioclase occurs in combination with quartz in plutonic rock fragments. These were not included in the plagioclase counts (see below). Patch and string perthites (Moorhouse, 1959), microcline and orthoclase are minor constituents. These occur

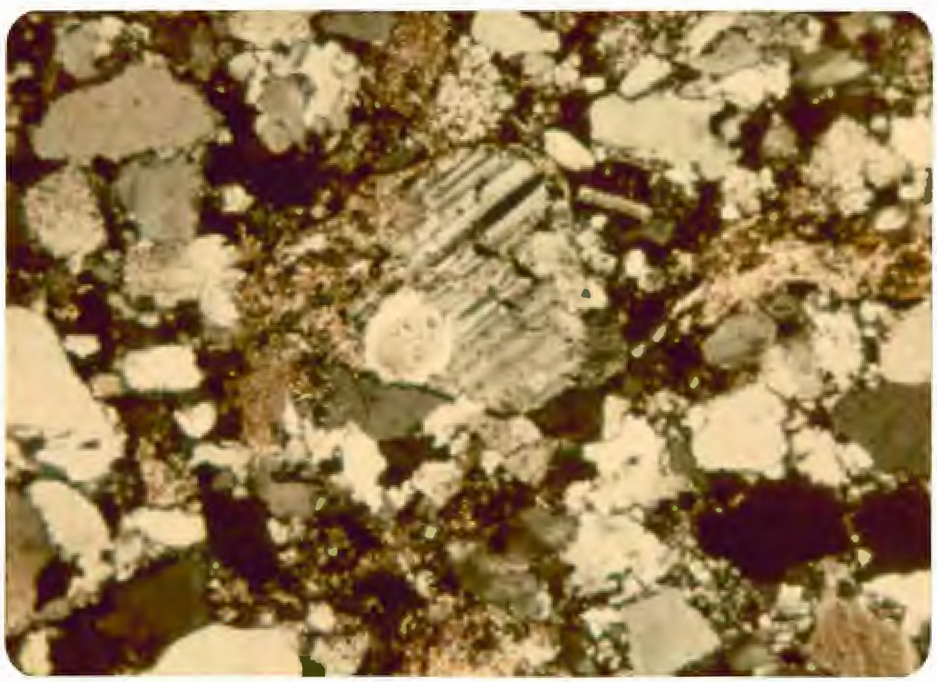


Figure 51--PHOTOMICROGRAPH OF GRAYWACKE WITH NORMALLY ZONED PLAGIOCLASE GRAIN (CENTER) WHICH HAS BEEN SOMEWHAT ALTERED TO SERICITE. Field of view approximately 2 mm. Crossed nicols.

as free grains and rarely in plutonic rock fragments. Small granitic veinlets which are primarily orthoclase occur in a few of the thin sections.

Volcanic rock fragments include felsic, intermediate and mafic varieties, with felsic types being most abundant. Both porphyritic (Figures 52 and 53) and nonporphyritic felsics are common. The nonporphyritic types are aphanitic, felty masses of feldspars (?) with a corroded, brown-stained appearance under plane-polarized light. Opaques, too finely divided to identify, are common inclusions. Under crossed nicols, extinction is segmented so that entire areas of the volcanic grain go extinct at once or, alternatively, so that extinction moves as a wave across segments of the grain. Boundaries between segments are highly irregular. Phenocrysts large enough to identify include quartz and rarely orthoclase.

In addition to the felted, corroded variety, other felsic volcanic rock fragments which are much cleaner appearing under plane-polarized light and exhibit a "wormy" texture with pinpoint extinction are found. These fragments are identifiable as volcanics when they are porphyritic (Figure 54) but are easily confused with coarsely crystalline chert when not. This problem has undoubtedly led to some misidentification.

There are some felsic volcanic rock fragments (chert?) which contain cubic opaques, both pyrite and pseudomorphic hematite after pyrite or magnetite. These grains could be iron formation fragments.

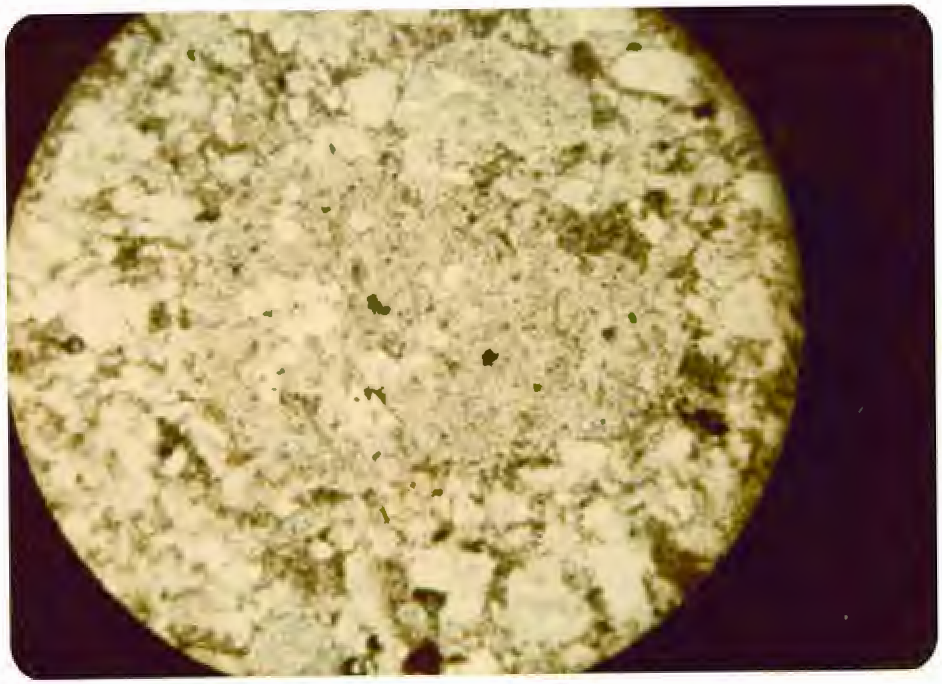


Figure 52--PHOTOMICROGRAPH OF GRAYWACKE WITH LARGE PORPHYRITIC FELSIC VOLCANIC ROCK FRAGMENT (CENTER). Field of view approximately 1. 6 mm. Plain light.

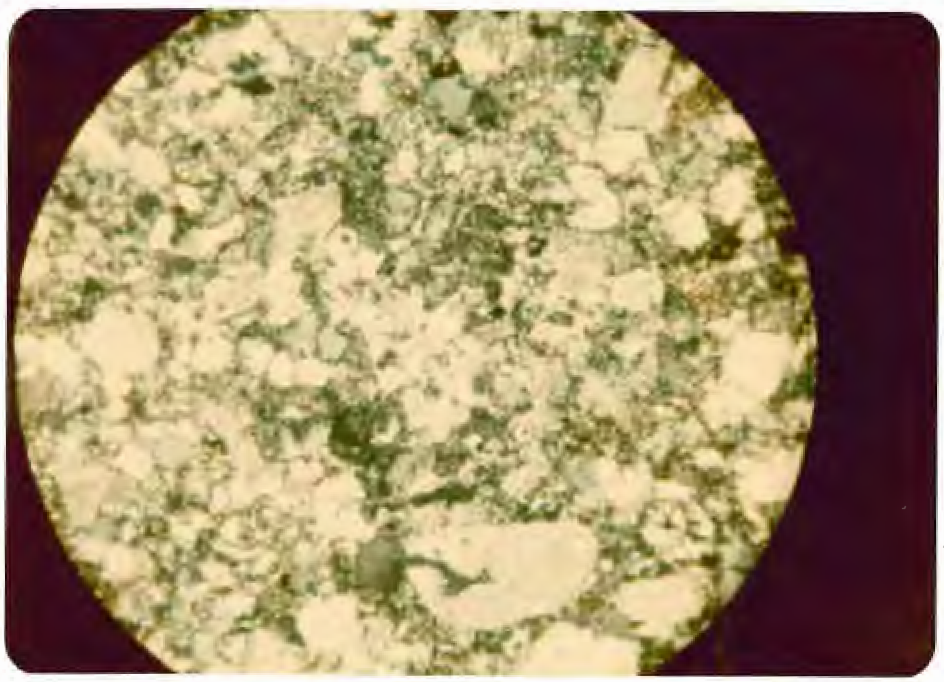


Figure 53--PHOTOMICROGRAPH OF SAME GRAYWACKE SHOWN IN FIGURE 52 (CROSSED NICOLS). Note how rock fragment "disappears" under crossed nicols.

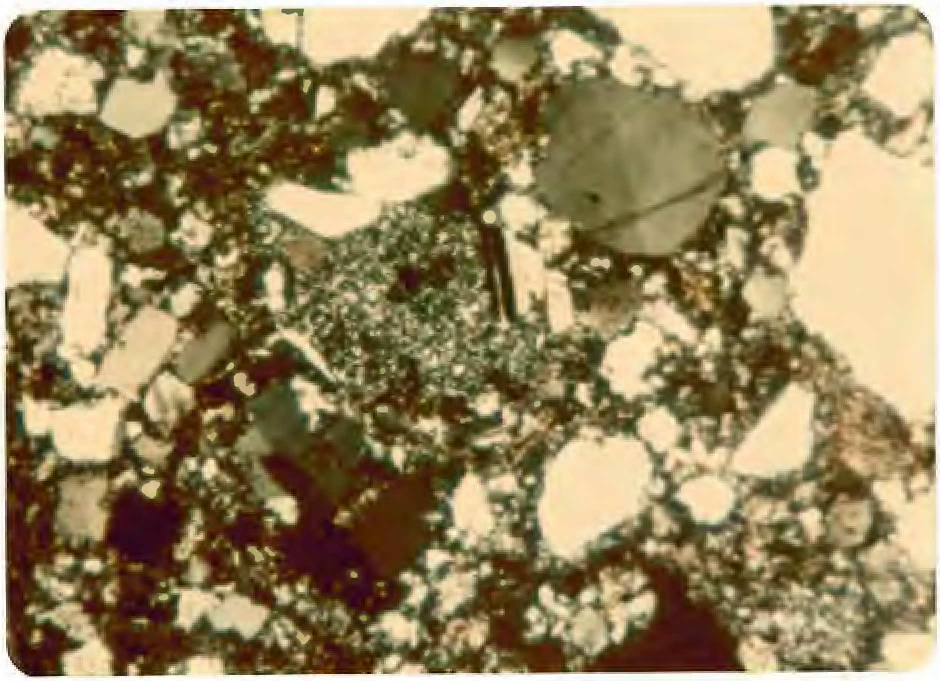


Figure 54--PHOTOMICROGRAPH OF GRAYWACKE SHOWING PORPHYRITIC FELSIC VOLCANIC ROCK FRAGMENT (CENTER) WHICH HAS A CHERT-LIKE, WORMY TEXTURE AND PINPOINT EXTINCTION. Phenocrysts are quartz and plagioclase. Field of view approximately 2 mm. Crossed nicols.

Alteration products of the felsic volcanics include sericite, biotite and chlorite. Fragments with completely altered ground masses are identifiable by the presence of unaltered quartz and feldspar phenocrysts or by the preservation of feldspar crystal outlines (oriented sericite) in a mass of unoriented alteration products.

Intermediate and mafic volcanic rock fragments (Figure 55) exhibit lath-shaped feldspars set in an aphanitic matrix which may be light colored (intermediate), dark or opaque (mafic). The groundmass is commonly altered to sericite and chlorite. All of the volcanic rock fragments are subrounded.

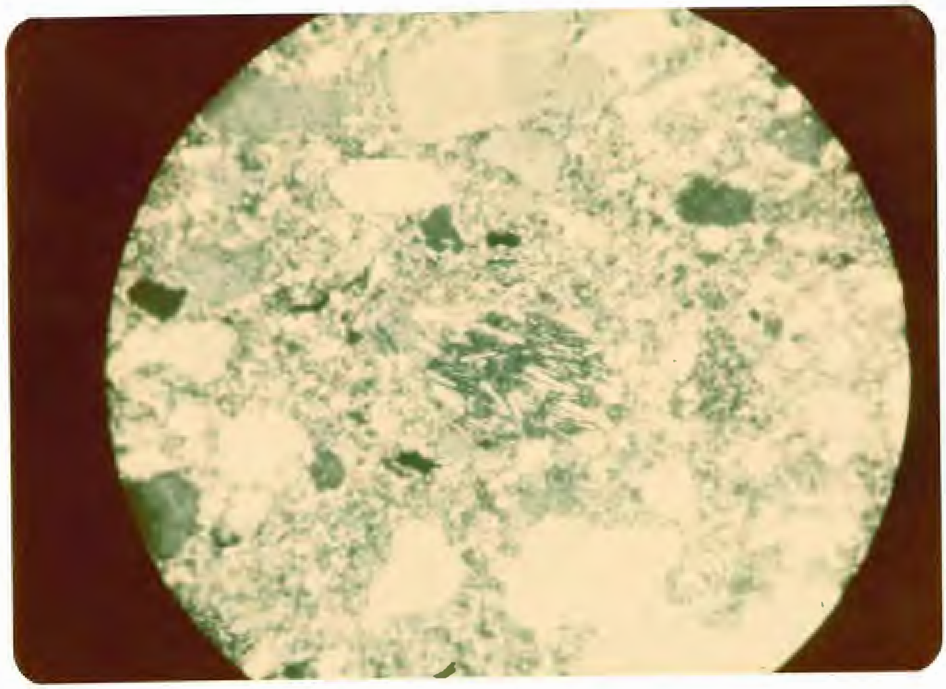


Figure 55--MAFIC VOLCANIC ROCK FRAGMENT (CENTER) CHARACTERIZED BY LATH-SHAPED FELDSPAR SET IN VERY DARK GROUNDMASS. Field of view approximately 1.6 mm. Crossed nicols.

Plutonic rock fragments are also found in the Tyler sandstones. Figure 56 shows an especially large granitic rock fragment. These fragments are most commonly quartz plus plagioclase suggesting an original granodioritic source rock, but some are comprised of quartz and orthoclase. Graphic intergrowths of quartz and feldspar were also noted and were included in the plutonic rock fragment count. Alteration of the feldspar in plutonic rock fragments is the same as that of free feldspar grains. Plutonic rock fragments are subrounded to subangular. Both volcanic and plutonic rock fragments are seldom larger than medium sand.

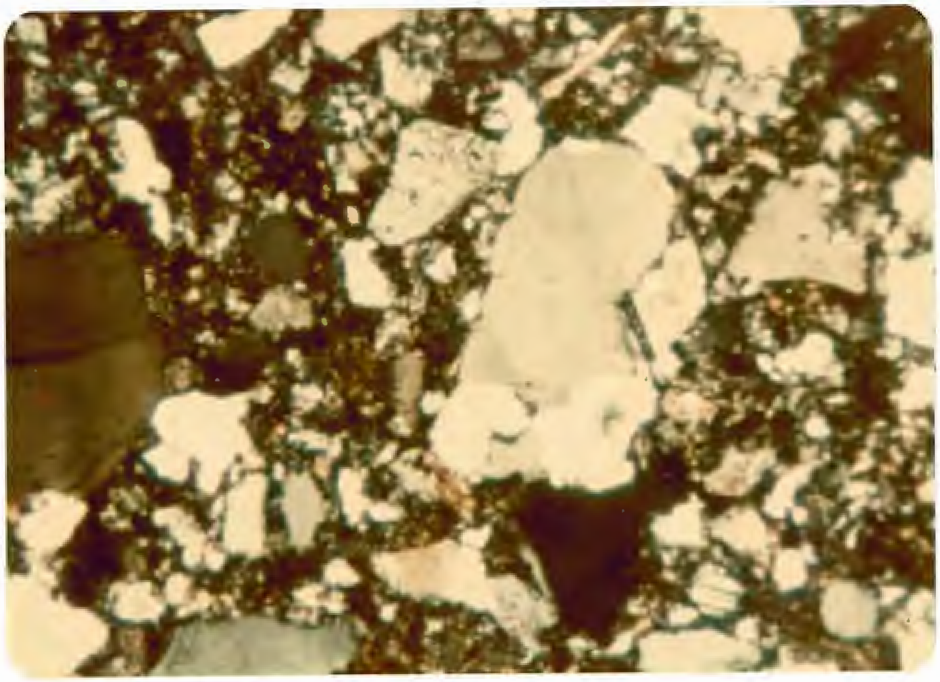


Figure 56--GRANITIC ROCK FRAGMENT IN GRAYWACKE.
Field of view approximately 2 mm. Crossed nicols.

Sedimentary rock fragments include lutite clasts, muddy siltstones, chert, chalcedony and rare reworked grains (see above). Lutite clasts are angular and blocky or wispy in appearance and characterized by much carbonaceous material. The carbonaceous matter occurs as opaque irregular stringers intermixed with a clayey matrix which is now partially or completely altered to sericite and chlorite. The lutite clasts are commonly brown or red brown in plane light and bent. The bending quite probably occurred during compaction. Carbonaceous muddy material squeezed around framework grains also suggests some post depositional deformation of these clasts. Lutite clasts are often silty and resemble muddy siltstone fragments. Lutite clasts can be extremely large (see <u>Description of Internal Bedding Features</u>). The muddy siltstone-silty mudstone break is set at the 50% silt level.

Muddy siltstone fragments are comprised primarily of very angular quartz grains in a muddy, now sericitized, matrix. Carbonaceous material is also present in these clasts. Both the muddy siltstone clasts and the lutite clasts may be of intraformational derivation. Distinction between altered muddy siltstones and porphyritic volcanic rock fragments whose matrices are also altered is difficult. If carbonaceous material is present they are counted as muddy siltstones and if feldspars are identifiable they are classified as volcanic; if neither criteria is usable, the grain in question is counted in the miscellaneous category (see below). Texturally, the siltstone fragments are very similar to the lutite clasts.

Chert and chalcedony are found as separate framework grains and rarely together in a single rock fragment. Pinpoint extinction and "wormy" texture distinguish chert from all other framework grains except felsic volcanics. In general this distinction can be made as well since the cherts are cleaner, without the corroded appearance in plane light and are equigranular whereas the felsic volcanics are often porphyritic. Partially recrystallized cherts may show areas of larger crystal size which could be construed as "porphyritic" but in all cases the large crystals grade into the cryptocrystalline groundmass. In porphyritic felsic volcanics there is no such gradation.

The cherts are generally more rounded than other framework quartz grains and sometimes have veinlets of larger quartz crystals, a probable result of diagenesis.

Chalcedony is identified by its fibrous, radial growth habit and the wavy extinction which accompanies it. Outlines of the radial habit are sometimes visible under plane-polarized light due to the presence of dusty inclusions. Chalcedonic fragments, unlike the cherts, are very angular. There are fragments of chalcedony with "wormy" crystal outlines and of chert with wavy extinction which probably are transitional phases between the two. Therefore, chert and chalcedony were counted together in the modal analyses. Both vary in size from silt to coarse sand.

A few sedimentary reworked quartz grains were found (Figure 50). They are monocrystalline quartz grains with optically continuous overgrowths. It is thought they are reworked, that is, derived from an older quartzose sandstone source, since they are set in a mud-rich matrix within which quartz overgrowths are unlikely to have grown. These grains are sand-sized.

Metamorphic rock fragments include argillite (slate?) clasts and stretched polycrystalline quartz clasts. The argillites have all the size, shape and mineralogical characteristics of lutite clasts with the added characteristic that the sericitic matrix shows a strong orientation. Whether the mica orientation is parallel to original bedding (an argillite fragment) or parallel to slaty cleavage (a slate fragment) is difficult to determine. But well-developed micas and recrystallization of the remainder of the fragment as well are suggestive of slate fragments. One such slate fragment with an obvious metamorphic texture complete with

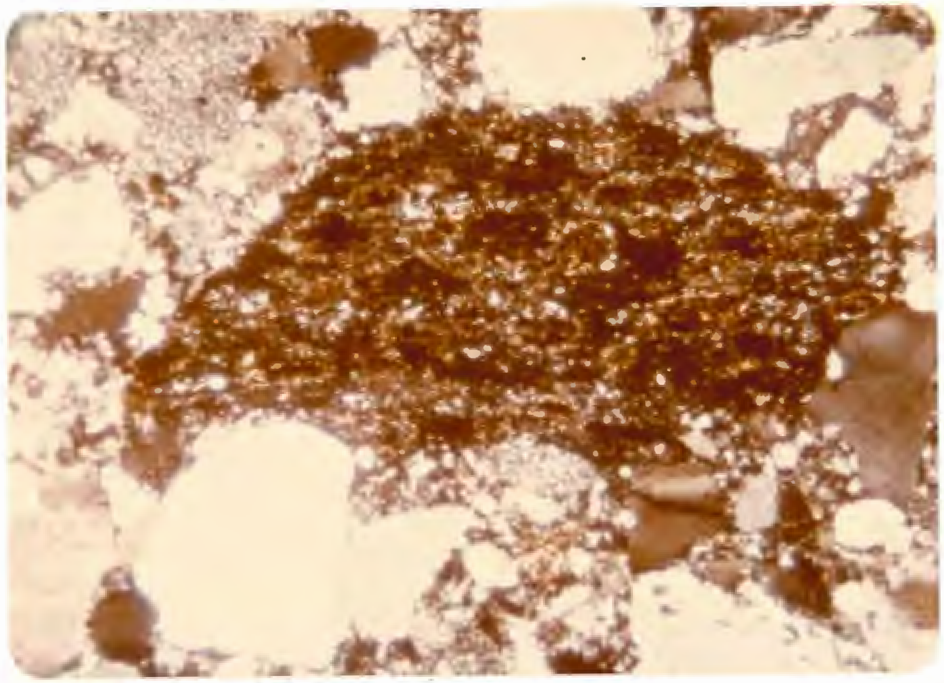


Figure 57--FRAGMENT OF "SPOTTED SLATE" IN COARSE GRAYWACKE. Compare texture of this fragment to Figure 58. Field of view approximately 2 mm. Crossed nicols.

porphyroblasts of cordierite (?) is shown in Figure 57.

Polycrystalline quartz grains characterized by elongate crystals are thought to be fragments of metamorphic rocks. Stretched metamorphic quartz of this type with included oriented micas are called schist fragments. Metamorphic quartz fragments are generally very angular to subangular, and are recognizable only in the sand fraction.

Opaques are found in all of the thin sections studied. They are either stringy carbonaceous materials or granular pyrite and hematite. It is difficult to determine with certainty if the metallics are original framework grains (detrital) or secondary but much is probably secondary.

Detrital micas include both muscovite and biotite but predominantly muscovite. Micas smaller than .03 mm. were considered matrix minerals, the result of recrystallization in the rock. Accessory framework minerals include zircon and green and brown tourmalines. A catch-all category was erected for point counting purposes for framework grains which were unidentifiable using the above definitions.

The slates and argillites were not particularly useful for petrographic study. Small grain size (when not metamorphosed) precluded compositional determinations. In the western end of the study area, the metamorphism has been such that cordierite (?) porphyroblasts have been developed in the muddy rocks ("spotted slates," see Figure 58). This metamorphism has obscured the original composition of the rocks and has made attempts at source area reconstruction even more difficult than it already is when dealing with such fine-grained rocks.

Modal analyses were performed on 27 of the graywackes of the Tyler Formation. A minimum of 600 points were counted on each thin section. Table 13 summarizes the results of those analyses. The first 17 samples in Table 13 are from the measured stratigraphic section (see STRATIGRAPHY AND LITHOLOGY) and are listed in correct sequence. That is, TS46 - 26 is the oldest and TS45 - 618 the youngest of the samples analyzed. As can be seen, there are no significant trends in composition of the Tyler graywackes up section. These subtle changes were noted in the first 17 samples: slight increase in percentage of monocrystalline quartz upward; slight decrease in percentage of polycrystalline quartz upward. There is also a more noticeable increase in the percentages of intermediate and mafic volcanic rock fragments up

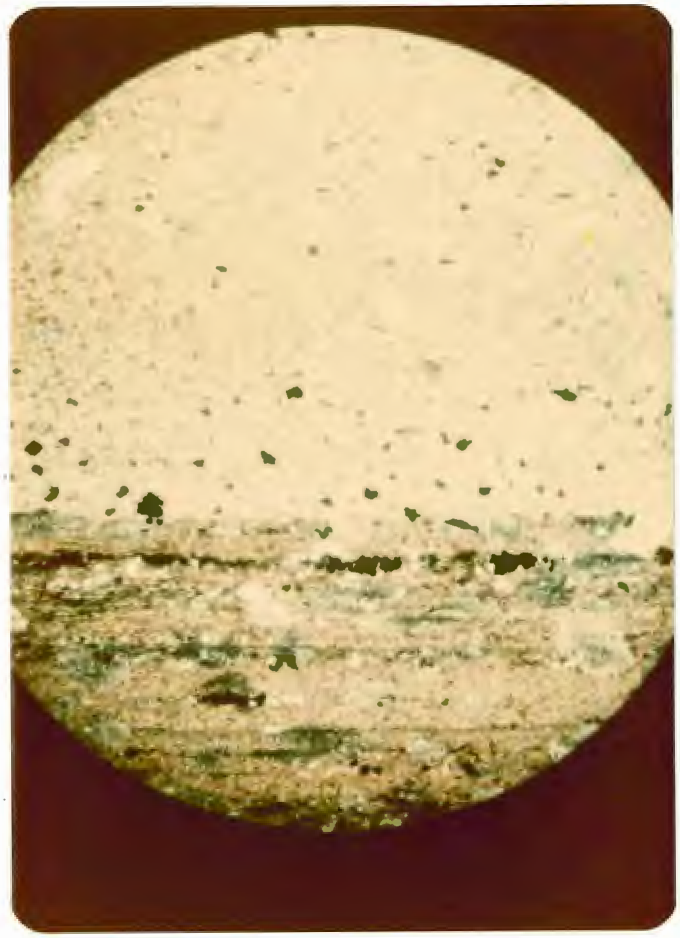


Figure 58--PHOTOMICROGRAPH OF SILT-STONE-SLATE BEDDING CONTACT. Note grading in siltstone and cordierite (?) porphyroblasts. Field of view approximately 6 mm. Crossed nicols.

section. No intermediate or mafic volcanic rock fragments were noted in the oldest samples while each of the youngest has at least one. The difference is slight, however, and to attach great significance to this trend would be prejudicial.

The last 10 samples of Table 13 were selected at random from the rest of the Tyler outcrop belt but are all from outcrops farther west

TABLE 13--MODAL ANALYSES OF TWENTY-SEVEN

Sample Number	Total Quartz			Total Feldspar				Total Rock Fragments							Opaques				
	mono. qtz.	poly. qtz.	chert + chalcedony	plagioclase*	orthoclase*	perthite	microcline	granitic	felsic volc.	intermediate + mafic volc.	metamorphic	sedimentary	greenstone	matrix	carbon. mat.	hematite	pyrite	chalcopyrite	leucoxene
TS45-618	22	22	1	6	tr	<1	tr	3	1	tr	tr	tr	tr	42	<1	tr	-	-	<1
TS45-487	12	38	2	4	2	tr	-	7	3	tr	-	1	-	28	1	-	-	-	-
TS45-397	22	21	2	1	2	tr	tr	5	3	<1	<1	15	tr	26	-	-	-	-	<1
TS45-380	19	25	1	3	3	tr	tr	4	4	tr	-	1	-	37	<1	1	tr	-	-
TS45-377	15	31	1	3	1	-	tr	3	3	tr	-	6	-	36	<1	-	-	-	-
TS45-243	16	31	2	5	0	tr	tr	6	7	-	tr	3	-	27	tr	-	-	-	-
TS45-241	11	26	1	4	0	tr	tr	5	4	tr	tr	15	-	31	<1	-	<1	-	-
TS45-147	23	21	<1	12	0	tr	-	4	6	-	tr	<1	-	32	-	-	-	<1	-
TS45-146	28	19	1	7	0	tr	-	2	5	-	tr	1	-	35	1	-	-	-	-
TS45-76	21	27	1	5	1	tr	-	4	3	-	-	7	tr	30	<1	tr	-	-	-
TS43B-78	17	26	1	8	<1	-	tr	2	1	tr	-	2	-	42	<1	tr	tr	-	tr
TS43-17	18	34	1	9	3	tr	tr	2	4	-	tr	tr	-	27	<1	-	-	-	tr
TS40-6	16	40	1	8	<1	-	-	3	2	-	-	-	-	26	†6	-	-	-	-
TS40-3	16	35	1	9	0	-	tr	1	1	-	-	-	-	35	2	-	-	-	-
TS46-109	17	28	3	6	3	tr	tr	4	1	-	-	tr	-	38	<1	tr	tr	-	tr
TS46-81	10	47	1	5	7	tr	tr	4	1	tr	-	1	-	23	2	-	tr	-	tr
TS46-26	13	21	<1	3	4	-	tr	1	<1	-	-	1	-	56	1	-	tr	-	-
TS8	3	66	-	2	3	-	-	4	-	-	-	-	-	22	<1	-	-	-	-
TS10A	.5	78	-	3	4.5	-	-	2.5	-	-	-	-	-	10	tr	-	-	-	-
TS14B	8	32	<1	6	1	-	-	10	8	-	tr	1	-	31	-	-	-	-	-
TS15	13	31	<1	5	3	tr	-	14	8	tr	-	1	-	23	-	-	<1	-	-
TS23B	7	33	1	6	tr	-	-	17	11	-	tr	1	-	22	1	-	-	-	-
TS24B	26	27	3	6	1	-	-	7	9	-	<1	2	-	17	<1	-	-	-	-
TS24M	21	41	1	5	1	tr	tr	4	9	-	<1	1	-	15	-	-	-	-	-
TS24T	9	58	<1	4	1	tr	tr	8	7	tr	-	<1	-	11	-	-	-	-	-
TS24C	18	34	2	4	1	<1	tr	6	4	-	-	11	-	19	-	-	-	-	<1
TS60	12	45	2	4	1	tr	-	7	6	-	-	1	-	22	-	<1	-	-	-
													average	28					

* orthoclase and plagioclase percentages adjusted to reflect visual esti-

<1 means at least 2 grains encountered during modal analysis

tr = trace = 1 grain

† included in matrix percentage

SAMPLES FROM THE TYLER FORMATION

zircon	tourmaline	detrital micas	mono. qtz. + overgrowths	mud(?)pellets	unknown	Recalculated to 100%			largest grain excluding mud frags. (mm)	Rock Name, after Pettijohn, Potter and Siever, (1973)
						quartz + chert	feldspar	rock fragments		
-	-	-	-	1	-	79	12	9	.5	feldspathic graywacke
-	tr	tr	tr	1	tr	76	8	16	.5	lithic graywacke
tr	-	-	-	1	<1	62	4	34	2	lithic graywacke
tr	tr	-	tr	1	-	75	11	14	1	lithic graywacke
-	-	tr	-	1	-	75	7	18	1	lithic graywacke
tr	-	-	1	1	tr	68	7	25	1	lithic graywacke
-	tr	-	-	tr	1	57	7	36	1	lithic graywacke
-	-	-	-	-	<1	66	17.5	16.5	1	feldspathic graywacke
tr	-	tr	-	tr	1	76	11.8	12.2	1	lithic graywacke
tr	-	tr	-	-	1	71	9	20	1.5	lithic graywacke
tr	-	tr	tr	tr	tr	77	14	9	.5	feldspathic graywacke
tr	-	-	-	<1	tr	75	17	8	.75	feldspathic graywacke
tr	tr	tr	-	-	tr	80	12	8	.25	feldspathic graywacke
tr	-	-	-	-	<1	83	13	4	.25	feldspathic graywacke
tr	tr	-	tr	-	tr	77	15	8	1	feldspathic graywacke
tr	-	-	tr	-	tr	77	15	8	.75	feldspathic graywacke
tr	-	tr	tr	-	tr	78	16	6	.5	feldspathic graywacke
-	-	-	-	-	-	88	7	5	.75	feldspathic graywacke
tr	-	-	-	-	-	88	9	3	.5	subarkose
tr	-	tr	-	-	-	61	11	28	.5	lithic graywacke
tr	-	tr	-	-	1	59	11	30	.75	lithic graywacke
tr	tr	tr	-	-	1	54	7	39	2	lithic graywacke
tr	-	tr	-	-	<1	68	8	24	1.5	lithic graywacke
tr	-	tr	-	-	1	76	7	17	1	lithic graywacke
tr	-	tr	-	-	tr	76	6	18	.75	sublitharenite
tr	-	-	-	-	-	68	7	25	2	lithic graywacke
tr	-	-	-	-	-	76	6	18	1.25	lithic graywacke
					average	73	10	17		

mates of stained slabs

than the first 17 samples. These samples appear to be somewhat richer in felsic volcanic and granitic detritus.

The "average" Tyler graywacke is 28% matrix with quartz and chert comprising 73% of its framework grains, feldspar 10% and rock fragments 17%. According to the classification system of Potter, Pettijohn and Siever (1973) it is a quartzose lithic graywacke. The very high percentages of quartz in all the samples warrants the "quartzose" modifier. Figure 59 is a plot of the modal analyses of the 27 selected samples. The apexes of the triangle are quartz plus chert (Q), feldspar (F) and rock fragments (RF). Just over half (15) of the samples plot as lithic graywackes. For comparison, the area where compositions of samples from presumably correlative formations, the Thomson (Morey and Ojakangas, 1970) and the Rove (Morey, 1969), would plot are also shown. Compositional plots of samples from correlative (?) formations in Michigan are not available.

From the preceeding discussion, the following conclusions are drawn. The terrain which was the source area of the Tyler sediments was primarily granitic in composition. Contributions from sedimentary, metamorphic and volcanic rocks were minor. There was also some contribution from the depocenter itself in the presence of unlithified mud clasts removed from areas upcurrent from their final resting places. Texturally, the sediments were very immature; typical of a rapid erosion-to-deposition history with little or no sorting.

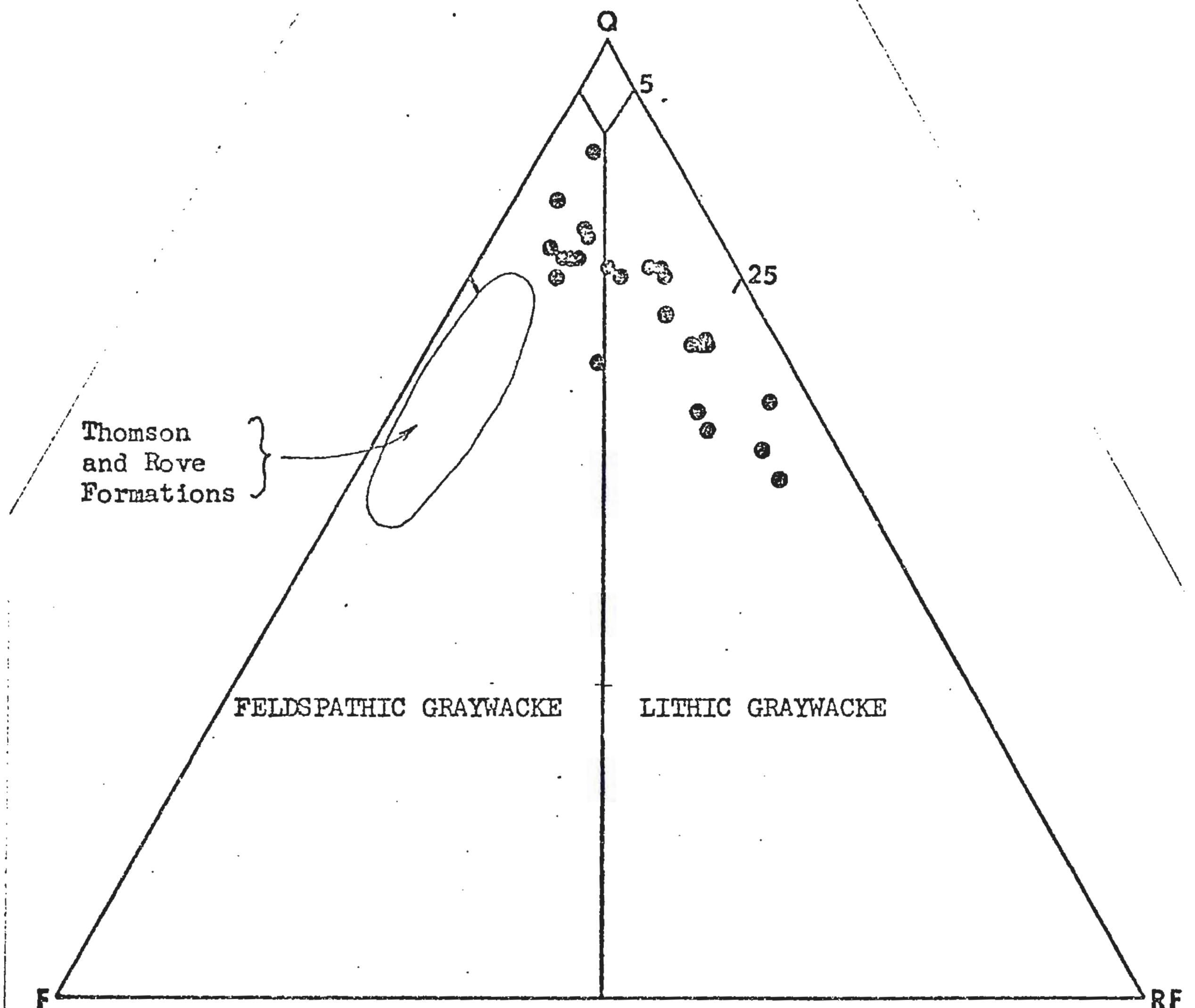


Figure 59--TERTIARY COMPOSITION PLOT OF TWENTY-SEVEN SAMPLES FROM THE TYLER FORMATION. Thomson (Morey & Ojakangas, 1970) and Rove (Morey, 1969) shown for comparison. Rock names from Pettijohn, Potter and Siever (1973). Compositions determined by modal analyses.

SEDIMENTATION

Provenance

Paleocurrent reconstruction and petrology both suggest that the source area was a highland which lay to the south and east and probably was much the same in character as that area is today. Lack of detailed mineralogic studies of the rocks which are today found in the presumed source area inhibits positive identification of the source. This study verifies the original suggestion by Irving and Van Hise (1883) that the Early Precambrian granites and granitic gneisses found to the south were the major source of detritus during deposition of the Tyler with minor contributions from the older greenstone terrain and possibly from the Early Middle Precambrian sedimentary rocks.

Both intrastratal and interstratal paleocurrent indicators point to a source which lay to the southeast of the present day position of the Tyler Formation. What minor variation which does occur in this generally consistent pattern is thought to be due to local submarine topography.

Tectonics and Sedimentation

The textural immaturity of the Tyler sediments suggests a relatively short time elapsed between freeing of the detrital grains from the parent rock and their final deposition, or that transport distance between source and deposition was short. Rapid erosion and high gradient are therefore called upon to remove the detritus and to quickly transport and deposit the sediments in the Tyler basin.

Although the sediments are compositionally submature (73% of the framework grains are quartz and chert), this maturity is thought to be partly a reflection of parent rock composition and not wholly the result of sedimentary differentiation. The angularity of the individual grains and the poorly sorted texture of the rocks support this interpretation. Relief between source and depocenter is thought, therefore, to have been moderate to great.

By inference, then, either the basin or the source, or both were tectonically active; the basin in a negative sense, the source positive. Thus, the environment was characterized by rapid sedimentation. Flysch-type sediments are commonly associated with early stages of orogenic activity. A change in the sedimentation rate (as documented in the thinning-upward trend of the argillaceous beds) suggests uplift in the basin probably slowly decreased and finally brought to an end the sedimentary episode which dominated Tyler time. Decreasing basin subsidence and finally elevation and exposure of the sediments to erosion marked the beginning of the Penokean Orogeny. Removal of the upper Tyler and possibly of succeeding formations by erosion which created the Middle Precambrian-Late Precambrian unconformity, makes interpretation of the exact nature of the change in tectonic environment speculative. Only the beginnings and end of that change are recorded in the Tyler sediments.

Within the basin area, the argillite and slate beds were almost certainly deposited during periods of relative quiescence. The sporadic intrusion of graywacke sediments was probably linked to earthquakes which probably accompanied orogenic activity and penecontemporaneous volcanism (to the east).

Current Mechanisms

It seems likely that the argillaceous beds were formed by a rain of pelagic sediments in a quiet water environment, periodically disturbed by strong currents. The lithologic associations, the assemblage of primary sedimentary structures and the microscopic character of the individual samples point most obviously to a turbidity current mechanism of transport and deposition of the coarser clastic beds. It should be noted, however, that a great many of the beds examined reveal none of the features indicative of turbidites. Other mechanisms, especially grain-flow, must be considered.

Environment of Deposition

By comparison of facies associations in the Tyler Formation to recent sediments and to ancient sediments of known depositional environments, it is concluded that the Tyler Formation was probably part of a submarine fan complex in fairly deep water. That complex periodically received influxes of sediment from shallower waters or directly from subaerial sources by way of a channel system. Once the currents that carried the sediments reached the bottom of the slope, they fanned out across the basin bottom, losing energy and dropping their sediments. The water at the site of deposition was both deep enough for pelagic sedimentation to have been the "norm" (i.e. below wave base) and near enough to shore to receive sporadic currents carrying coarser sediments.

Basin Geometry

The Tyler Formation was probably deposited in an intracratonic basin which was elongate, northeast to southwest. The basin was at least as long as the present Tyler outcrop belt (60 miles) and probably extended eastward to include the Copps and, possibly, Michigamme Formations. Goldich, et al. (1961) suggest the basin may have been continuous into the Mistassini District of Canada, some 900 miles to the northeast. Dry land, which was the Early Precambrian terrain, bounded the basin on the south and southeast and the shoreline probably lay close to the present Early Precambrian-Middle Precambrian contact.

The extent of the depocenter to the northwest is unknown. Tentative correlation of the Tyler with the Rove and Virginia Formations of Minnesota and Ontario suggest a basin width of at least 185 kilometers. However, Chase and Gilmer (1973), and others, suggest that the Midcontinent Gravity High, a geophysical and geologic feature which extends from the western Lake Superior basin to Abilene, Kansas, was the site of aborted Keweenawan rifting between Precambrian Plates. According to their calculations a 75-80 kilometer wide rift zone opened in Keweenawan time between the Mesabi and Gunflint ranges, where the Virginia and Rove Formations crop out, and the Gogebic Range. Thus, the original basin, if the Tyler, Virginia and Rove all were deposited in a common basin, was probably closer to 105-110 kilometers in width.

Reconstruction of a paleoslope perpendicular to paleocurrent trends in the Tyler, Rove and Thomson (Virginia?) Formations (Figure 60) supports the idea of a single Middle Precambrian sedimentary basin bounded on the north-northwest and south-southeast by crystalline highlands. These highlands were the source for the sediments (Morey, 1969; Morey and Ojakangas, 1970; this paper) which were deposited in a subsiding, deep water basin.

It is thought, therefore, that the Tyler, Virginia (Thomson), Rove, Copps, Michigamme and probably Rabbit Lake (Cuyuna Range) Formations occupied different loci of a common depositional basin. The basin was landlocked on at least three sides, but may have been open to the northeast.

Table 14 is a sedimentary model for the Tyler Formation and correlative (?) formations of the Lake Superior area.

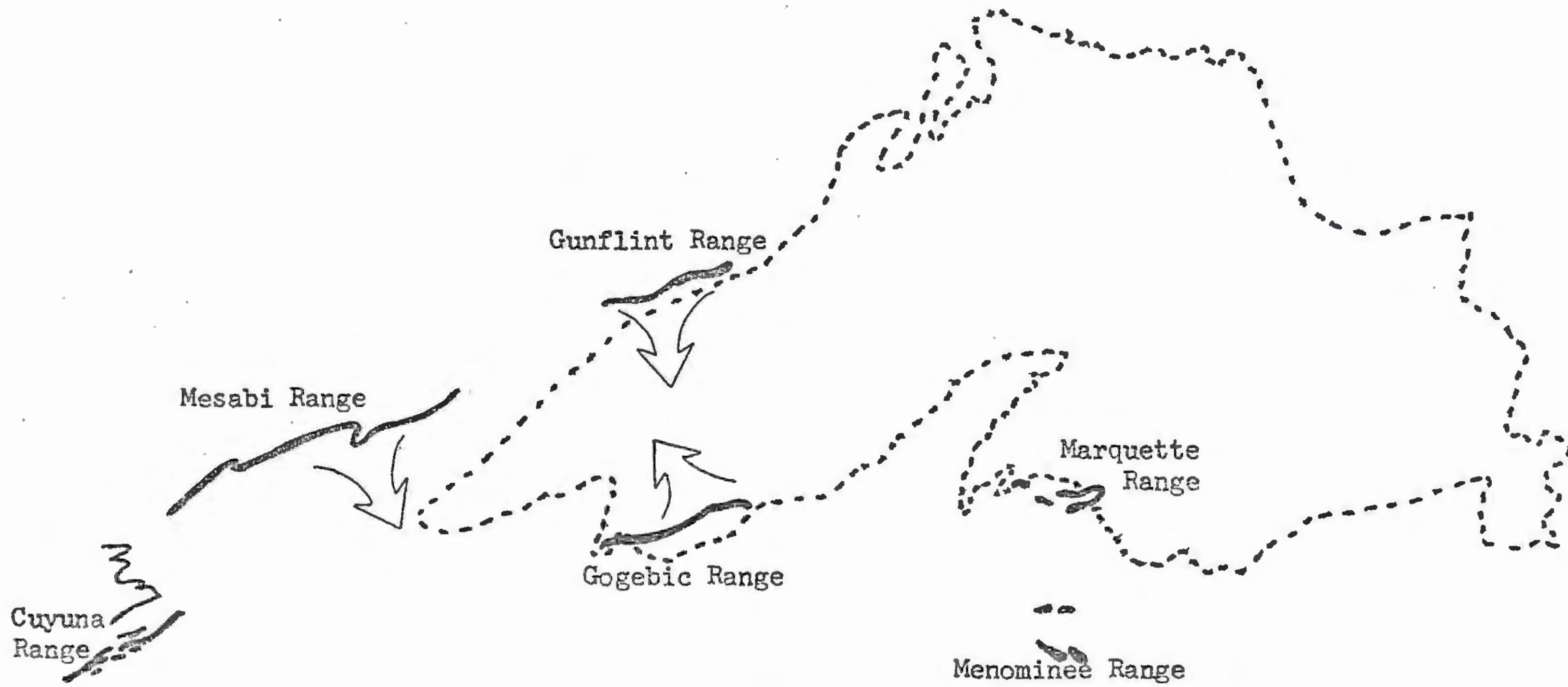


Figure 60 --PALEOCURRENT TRENDS IN MIDDLE PRECAMBRIAN SEDIMENTS OF THE LAKE SUPERIOR REGION. Iron Ranges have been restored to pre-rift positions (see text). Outline of Lake Superior shown for reference. Paleocurrent trends shown for the Rove of the Gunflint Range, the Thomson (Virginia?) of the Mesabi Range and the Tyler of the Gogebic Range. The figure points out the need for paleocurrent studies in the Rabbit Lake (Cuyuna Range) and Michigamme (Marquette Range) Formations.

TABLE 14--SEDIMENTARY MODELS FOR THE TYLER FORMATION AND OTHER FORMATIONS OF MIDDLE PRECAMBRIAN AGE. (after Morey and Ojakangas, 1970)

References / Attribute	Tyler Formation (this paper)	Thomson Formation (Morey & Ojakangas, '70)	Rose Formation (Morey, 1969)	Virginia Formation (Pfleider & others, 1968, White, 1954)	Rabbit Lake Formation (Schmidt, 1963)	Michigamme Slate (James, 1958; Nilsen, 1965)
Basin Geometry	Basin elongated east-west parallel with tectonic strike; regional paleoslope dips toward north; loci of maximum accumulation unknown.	Basin elongated east-west parallel with tectonic strike; regional paleo slope dips toward south; loci of maximum accumulation unknown.	Basin elongated east-west parallel with tectonic strike, paleoslope dips toward the south, loci of maximum accumulation probably to the south.	Basin elongated east-west parallel with tectonic strike, paleoslope probably dips toward the south, loci of maximum accumulation probably to the south.	Basin probably elongated east-west parallel with tectonic strike; shoreline north-south with paleoslope dipping to the east; loci of maximum accumulation probably to the east.	Basin elongated east-west parallel with tectonic strike, paleoslope unknown, loci of maximum accumulation unknown.
Directional Structures	Small-scale cross-stratification, flute and groove casts, negative grooves (ridge molds); all (mean 290°) indicate paleoslope and are normal to depositional strike; solemarks and cross-stratification in general agreement.	Chiefly small scale cross-stratification, the dip azimuth (mean 176°) indicates the paleoslope and is normal to the depositional strike; flute and groove casts indicate southward-trending currents but have a wide scatter.	Small-scale cross-stratification, flute and groove casts, all (mean 170°) indicate paleoslope and are normal to depositional strike, ripple marks indicate southward flowing longshore currents.	Cross-stratification, but direction unknown.	Unknown.	Mostly unknown, cross stratification indicates current flow to the southeast.
Lithic Fill	Interbedded, graded and nongraded feldspathic and lithic graywacke, siltstone and slate (argillite); rare carbonate concretions; ferruginous near base.	Interbedded, graded feldspathic graywacke, siltstone and slate, abundant carbonate concretions.	Interbedded graded feldspathic graywacke siltstone and argillite, abundant carbonate concretions, chert lenses, especially near base; siltstone and graywacke more abundant upward in section, minor quartzite in upper 700 feet.	Interbedded graded graywacke, siltstone and slate, abundant carbonate concretions & chert lenses near base, siltstone and graywacke more abundant upward in section.	Interbedded graywacke, siltstone and slate, slate apparently most abundant; lava flows(?) locally near base; interbedded lenses of iron-formation.	Interbedded graded quartzose and feldspathic graywacke and slate, scattered concretions and local quartzite units.
Thickness	At least 12,000 feet.	At least 3,000 feet and perhaps as much as 20,000 feet.	0-to at least 3,100 feet.	0-to at least 3,000 feet.	At least 2,000 feet.	Minimum of about 5,000 feet to a possible maximum of about 11,000 feet.
Arrangement	Sand-shale ratio increases upward; marginal facies removed by pre-Keweenawan erosion.	Unknown.	Sand-shale ratio increases upward; marginal facies removed by pre-Keweenawan erosion.	Sand-shale ratio increases upward; marginal facies removed by pre-Keweenawan erosion.	Unknown.	Unknown.
Tectonic Setting	Probably miogeosynclinal-moderately negative.	Unknown; probably miogeosynclinal – moderately negative.	Miogeosynclinal - moderately negative.	Miogeosynclinal – moderately negative.	Unknown.	Strongly negative – abundant associated volcanic rocks, etc.

CONCLUSIONS

1. The name Tyler Slates is a misnomer. Sandstone and siltstone are volumetrically more abundant (59%) than mudstone (41%). The name Tyler Formation is therefore preferable.

2. Subdivision of the Tyler Formation into separate members based on lithologies recognizable in the field is impossible. Seven outcrops near Hurley, Wisconsin give the best representative or type section of the Tyler.

3. A decreasing thickness trend in the argillaceous beds suggests a decreasing rate of pelagic sedimentation upsection which could have been the result of decreased subsidence and subsequent uplift in the basin heralding the beginning of the Penokean Orogeny.

4. Primary sedimentary structures suggest transport and deposition of the coarse-grained Tyler sediments by turbidity currents and, possibly, grain-flow mechanisms.

5. Paleocurrent indicators show that the turbidity currents flowed north and northwest down the paleoslope at velocities of a few tens of centimeters per second.

6. The Early Precambrian granite and granite gneiss terrain to the south appears to have been the major source area of the sediments. Minor contributions came from felsic volcanics and older,

metamorphosed sediments.

7. The Tyler Formation was deposited in relatively deep water as a submarine fan complex which had an original slope of approximately $0^{\circ}10'$.

8. The Tyler Formation and the presumably correlative formations of the Lake Superior region (the Rove, Virginia, Thomson, Rabbit Lake, Copps and Michigamme Formations) may have been deposited in a common intracratonic basin. Studies to date suggest different source areas and different paleocurrent orientations. The coarse clastic rocks of the Rove and Thomson Formations are feldspathic graywackes while those of the Tyler are quartzose lithic graywackes.

FIELD WORKSHEET - TURBIDITES

rmation

: #

ation

from strat. top or bot.

itude ______ ______ same
strike dip

ty cleavage ______ ______ same
strike dip

	min	max	are the contacts: gradational sharp
T_e			
T_d			
T_c			
T_b			
T_a			
sole marks?			
total	______		covered interval?

a Units

unit: flames or loaded flames ______ grn size ______
orientation (if other than mud)

unit: thkness of indiv. laminae ______ grn size ______

unit: grn size ______
X-bedding: type ______ orientation ______ ampl. ______ ______
convoluted: axis orientation ______ ______ truncated crests? yes no can't tell
plunge strike

unit: thkness of indiv. laminae ______ grn size ______

unit: grn size ______ lgst grn size ______ nongraded or graded:
or ______ grades to ______ shale clasts? yes ______ ______ ______
no length width height

internal sed. features:
1. laminated
   thkness of indiv. laminae ______
   thkness of lam. zone ______
2. rippled: straight, caternary, linguoid, lunate, cuspate, interference, can't tell
3. X-bedded: parallel, wedge, parallel trough, festoon, can't tell
   orientation______ ampl.______ ______
4. other: Lutite clasts abundant rare largest clast ______

nongraded or graded:
1. ideal
2. graded base only
3. graded top only
4. multiple ______ X
5. laminated graded
6. reverse
7. apparent reverse
8.

MARKS

	orientation	direction	splayed	sym.	length	width	relief
ove casts							
te casts							
d marks							
er ______							

APPENDIX B

AVAILABLE THIN SECTIONS OF ROCKS FROM THE TYLER FORMATION

At the University of Wisconsin-Madison Department of Geology Repository:

Tabet (1974)

UW 1607-66	hornblende hornfels
-67	feldspar-quartz-hornfels
-68	hornfels (Tyler?)
-69	quartz-feldspar-hornfels
-70	cordierite schist
-71	cordierite schist

Cooper (1973)

UW 1579-43	quartz-rich graywacke
-44	graywacke
-45	shale with secondary pyrite

Felmlee (1970)

UW 1541-58	Spotted slate
-59	cross-laminated siltstone
-60	graywacke

Some thin sections also at the Wisconsin Geological and Natural History Survey.

Thin sections from this study at University of Minnesota-Duluth Department of Geology.

APPENDIX C

CALCULATION OF CURRENT VELOCITY BY LARGEST CLAST METHOD (KOMAR, 1970)

Komar (1970) used cobble size to estimate competency of a turbulent current. He calculated the minimum critical shear stress necessary to entrain and suspend near-spherical cobbles and the minimum velocity necessary to create that shear stress. His cobbles exhibited a high degree of sphericity, so he did not need a shape factor in his computations. The same computations have been made for a large lutite clast from the Tyler Formation and provision has been made for the hydraulic effects of clast shape.

The largest lutite clast for which three dimensional measurements were made is 30 cm. by 13 cm. by 2.5 cm. (much larger clasts showing only two dimensions were found). By assuming the general shape of the individual lutite clast to be a triaxial ellipsoid (observation of clasts in outcrop suggest this geometry), then the volume of the large lutite clast under consideration is:

$$\text{volume} = \frac{\pi}{6} (LIS) \qquad (1)$$

where L, I and S are the longest, intermediate and shortest axes of the ellipsoid. The volume of the clast is, by equation (1), 510 $cm.^3$ The nominal sphere (a sphere with the same volume as the clast) has a diameter (d) of 9.9 cm. from the formula for the volume of a sphere:

$$d = \sqrt[3]{\frac{vol}{\frac{1}{6}\pi}} \quad (2)$$

Now, by using Komar's equation:

$$\tau_o = 0.06\,(\rho_s - \rho_t)\,g\,D \quad (3)$$

where: τ_o = critical shear stress,

ρ_s = density of the clast = approx. 1.40 gm cm^{-3} (Dott, 1972),

ρ_t = density of the current = approx. 1.18 gm cm^{-3} (Komar, 1970),

g = acceleration due to gravity = 980 cm sec^{-2}, and

D = diameter of the clast = 9.9 cm (equation 2),

the shear stress necessary to entrain and transport a spherical clast of the stated diameter is found to be 128 dynes. By Komar's equation relating critical shear stress and velocity:

$$\bar{u} = \left(\frac{\tau_o}{C_f\,\rho_t}\right)^{1/2} \quad (4)$$

where: $\bar{u}$ = velocity necessary to create the critical shear stress τ_o, and

C_f = drag coefficient = approx. .0035 (Komar, 1970),[1]

the estimated velocity necessary to support a spherical clast of the stated volume is found to be 176 cm. sec.$^{-1}$ But the clast is not spherical, so a smaller velocity is necessary. This velocity can be estimated by use of

[1] There is disagreement on the value of C_f between authors. See Hand, 1974, 1975 and Komar, 1970, 1975.

equation (5) which comes from Sneed and Folk (1958):

$$\psi_p = \sqrt[3]{\frac{S^2}{LI}} \qquad (5)$$

where: ψ_p = maximum projection sphericity = effective settling velocity (Blatt, Middleton and Murray, 1972).

By equation (5), effective settling velocity of the ellipsoid of the stated dimensions is .252 or 25% of the settling velocity of a nominal sphere. Alternatively, the velocity of the fluid medium necessary to support that same clast is 25% of the velocity necessary to support the nominal sphere. Therefore, the minimum velocity of the current which transported the clast described is estimated to be:

$$\text{Vel.} = (.25)\ (176 \text{ cm sec}^{-1}) = 44 \text{ cm sec.}^{-1}$$

It is suggested that these calculations yield only "ball park" velocity figures. The exact values of the parameters for which estimates were made (e.g. density of the turbulent current and density of the unlithified lutite clast) are unknown.

APPENDIX D

CALCULATION OF CURRENT VELOCITY AND PRIMARY SLOPE FROM RIDGE MOLD SPACING

Let T be the transverse spacing of the head lobes (interlobe width of Figure 44). At most, T would be equal to the entire width of the turbidity current head and at least, T would be equal to zero. The former would be a turbidity current head without lobes (riding on a cushion of ambient water) and the latter a current head without tunnels (the head everywhere in contact with the bedding plane). These two extremes are not practical (Allen, 1971b). A more reasonable estimation would be for interlobe spacing (T) to approximate lobe width (L of Figure 44). Logic dictates that due to the relative incompressibility of water, the volume of ambient water displaced by an encroaching lobe must be equal to the volume of ambient water flowing through a tunnel unless some water is pushed ahead of the turbidity current. Allen (1971b) rejects this latter alternative because empirical observation does not support an induced current ahead of an approaching turbidity current. Therefore, assuming regular spacing of lobes and tunnels (Allen, 1971b; Simpson, 1969), the interridge spacing (R) should equal the interlobe spacing (T) plus the lobe width (L):

$$R = L + T \tag{7}$$

and since by assumption,

$$L = T \tag{8}$$

then, $$R = 2T \quad (9)$$

or, $$T = R/2 \quad (10)$$

In turn, by Allen's (1971b) relationship:

$$T = d(c) \quad (11)$$

where: d = height of the turbidity current head, and

c is a constant which varies from .2 to .5 (Allen, 1971b, Simpson, 1969); and by rewriting equation (11), and by substituting for T from equation (10)

$$d = T/c = R/2(c) \quad (12)$$

Since c varies from .2 to .5, then:

$$d_{max} = R/.4 \quad (13)$$

and $$d_{min} = R/1.0 \quad (14)$$

where: d_{max} and d_{min} are the maximum and minimum head heights of a turbidity current with an interridge spacing of R. Using an average value of R = 3 cm. (measured in the field), d_{max} becomes 7.5 cm. and d_{min} is 3 cm.

These computations are of course, crude and based upon assumption. There is also some question of the validity of such calculations when dealing with spreading flow since the equations here cited were derived from flume experiments and restricted flow. It is suggested, however, that the resultant figures can be used as an "order of magnitude" approximation for turbidity current size. It is further suggested that negative grooves are characteristic of flows which do not have the necessary internal turbulence (i.e. velocity) and/or the necessary competency to create flute marks. It is this lack of strong internal turbulence which allowed the ridges to survive. In fact, Dzulynski and Walton (1965) experimentally found a gradational relationship (in terms of velocity) between flute marks and longitudinal ridges.

It is now possible to estimate the velocity of the turbidity current because the head height is related to velocity by the equation (Middleton and Hampton, 1973):

$$V = (0.7)\sqrt{\left(\frac{\Delta \rho}{\rho_a}\right) g d_2} \qquad (15)$$

where: ρ_a = density of the ambient water ($\approx$ 1.05 grams per centimeter^3 for salt water),

$\Delta \rho$ = difference in the densities of the turbidity current ($\approx$1.18, Allen 1971a) and the ambient water,

g = 980 cm. sec^{-2},

$d_2 = d_{max}$ = 7.5 cm.,

V = velocity of the turbidity current.

By equation (15), the velocity of the turbidity current which built the ridges spaced 3 cm. apart is calculated to be 21 cm. sec^{-1}. This velocity is of the same order of magnitude as that calculated by different methods in Appendix C.

Further, by Komar's (1970) equation relating critical shear stress (τ_o) and velocity (U) (equation 4, this paper), and by solving equation (4) for τ_o:

$$\tau_o = (U)^2 (C_f) (\rho_t) \qquad (16)$$

and since $U \approx V$,

then, $$\tau_o = (V)^2 \; (C_f \rho_t) \qquad (17)$$

where: τ_o, V and C_f are as above and ρ_t = density of the turbidity current ≈ 1.18 gmcm^{-3}. Therefore, by equation (17) and the values of V computed from equation (15), $\tau_o = 1.82$ dynes. Then by Komar's (1970) equation:

$$\tau_o = (\rho_t - \rho_a) \frac{gh \sin\beta}{1.5} \qquad (18)$$

where, β = slope of basin bottom.

Solving for $\sin\beta$:

$$\sin\beta = \frac{1.5\,\tau_o}{(\rho_t - \rho_a)\, gh} \qquad (19)$$

By using the value of τ_o as computed from equation (17) and by using the same values for ρ_t, ρ_a, g and h as used elsewhere, $\sin\beta$ is computed to be .002857 from equation (19). Therefore the primary slope is arc $\sin\beta$, or $.1636^{\circ}$. This value ($0^{\circ}9.8'$) falls within the range of slopes observed on modern submarine fans (see Table 12).

APPENDIX E

Curray's (1956) vector analysis of paleocurrent indicators as modified by Dott (1972) was chosen for analysis of indicators in the Tyler Formation. The equations used in the analyses of unipolar indicators are:

$$\tan \overline{\theta} = \frac{\Sigma \sin \theta}{\Sigma \cos \theta} \tag{20}$$

where, $\overline{\theta}$ = azimuth of the mean vector,
θ = azimuth of the individual vectors,

$$\text{and, } \quad r = \sqrt{\left(\frac{\Sigma \sin \theta}{N}\right)^2 + \left(\frac{\Sigma \cos \theta}{N}\right)^2} \tag{21}$$

where, r = magnitude of the resultant mean vector and r varies from 0 to 1.0;
$r = 0$ means complete randomness (i.e. maximum dispersion),
$r = 1$ means perfect orientation of all indicators.
(r may be expressed as a percentage by multiplying by 100)
N = the number of indicators.

The equations used in the analyses of bipolar indicators are:

$$\tan 2\overline{\theta} = \frac{\Sigma \sin 2\theta}{\Sigma \cos 2\theta} \tag{22}$$

where $\overline{\theta}$ and θ are as above, and,

$$r = \sqrt{\left(\frac{\Sigma \sin 2\theta}{N}\right)^2 + \left(\frac{\Sigma \cos 2\theta}{N}\right)^2} \tag{23}$$

where r, 0 and N are as above.

The last step in analysis of both unipolar and bipolar data is visual inspection of the plotted data (Figures 43 and 45) to eliminate unreasonable vector solutions yielded by trigonometric functions.

Vector analyses were performed separately for all of the following: flute casts, flame structures, cross-bedding, groove casts, negative grooves and parting lineations. The resultant vectors are plotted in Figures 43 and 45.

Angular deviation (S) was then computed for each of those six groups so that an F test of significance could be performed. The formula used was:

$$S = \sqrt{2(1-r)} \left(\frac{360^{\circ}}{2\pi}\right) \tag{24}$$

These calculations gave S values in degrees which were then used to calculate the statistical parameter F by the equation:

$$F = \frac{10,800}{S^2} \tag{25}$$

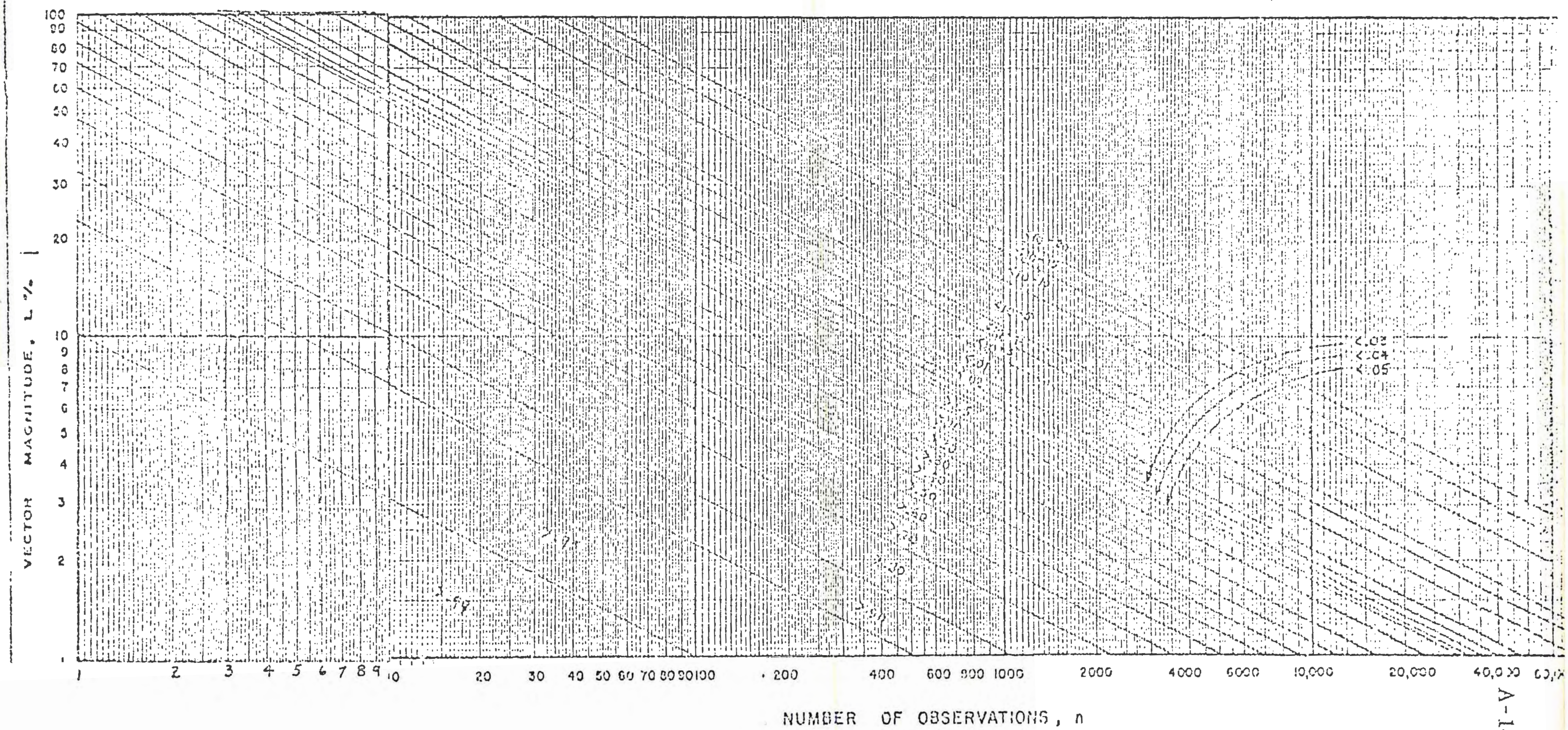
These resultant values of F for each of the six groups were then compared to F values in a standard F table (Croxton and Cowden, 1956) at the .05 significance level at the appropriate degrees of freedom (N-1). If the calculated value for the .05 significance level exceeds that value given in the F table, the orientation suggested by the paleocurrent measurements is significantly different from a chance orientation due to random distribution and the vector mean was considered reliable. The results of the F test are given in Table E-1.

A graphical test of significance was also performed for each of the six groups. The graphical test is that of Curray (1956) as modified

TABLE E-1--STATISTICAL TESTS OF SIGNIFICANCE - F TEST

indicator	F	N	N - 1	degrees of freedom	F table values at significance level:		significant (?)
					.05	.01	
flute casts	4.568	10	9	8	3.072	5.057	yes
flame structures	2.675	3	2	2	9.552	30.817	no
cross-bedding	3.146	31	30	30	1.84	2.39	yes
negative grooves	3.042	16	15	12	2.425	3.553	yes
groove casts	2.106	24	23	24	1.984	2.659	yes
parting lineations	7.054	6	5	4	4.534	9.148	yes

slightly to accommodate the low number of observations for some of the paleocurrent indicators in the Tyler Formation. The modified graph will be found on page A-14. The .05 level is commonly the dividing line between nonsignificant ($>$ .05) and significant ($\leq$ .05) data (Dott, 1972). The results of the graphical test of significance are given in Table E-2. From the results of these tests it is considered that flute casts, cross-bedding, negative grooves, groove casts and parting lineations yield statistically valid vector means, but that flame structures do not. Graphical tests of significance were also performed for each of the six geographic areas. The results are shown in Table E-3.



CURRAY, 1956
—Rayleigh test of significance
MODIFIED BY ALWIN, 1975

TABLE E-2--GRAPHICAL TEST OF SIGNIFICANCE

Indicator	Vector magnitude (%)	N	Significance[1]
flute casts	64	10	2/100, yes
flame structures	38	3	60/100, no
cross-bedding	48	31	<1/1000, yes
negative grooves	46	16	<4/100, yes
groove casts	22	24	<30/100, no
parting lineations	77	6	4/100, yes

[1]Less than .05 is significant
Greater than .05 is not significant

TABLE E-3--TESTS OF SIGNIFICANCE FOR GEOGRAPHIC AREAS AND INDICATORS

Area and indicators	F	N	N - 1	degrees of freedom	F table values at significance level: .05	F table values at significance level: .01	significant (?)
Subdivision 1							
unipolar	2.188	21	22	24	2.054	2.801	yes
bipolar	1.767	8	7	6	3.581	6.371	no
Subdivision 2							
unipolar	6.143	4	3	3	6.591	16.694	no
bipolar	2.030	5	4	4	5.192	11.392	no
Subdivision 3							
unipolar	301.	2	1	1	18.513	4052.2	yes
bipolar	2.852	13	12	12	2.604	3.960	yes
Subdivision 4							
unipolar	none	none	none	none	none	none	none
bipolar	8.565	5	4	4	5.192	11.392	yes
Subdivision 5							
unipolar	7.082	7	6	6	3.866	7.191	yes
bipolar	3.542	9	8	8	3.230	5.467	yes
Subdivision 6							
unipolar	3.555	2	1	1	18.513	98.503	no
bipolar	5.581	7	6	6	3.866	7.191	yes

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