1988

Stratigraphy and sedimentation of the Ravalli Group (middle Proterozoic Belt Supergroup) in the Mission Swan and Flathead Ranges northwest Montana

Christopher Cronin
The University of Montana

STRATIGRAPHY AND SEDIMENTATION OF THE RAVALLI GROUP (MIDDLE PROTEROZOIC BELT SUPERGROUP) IN THE MISSION, SWAN, AND FLATHEAD RANGES, NORTHWEST MONTANA

by

Christopher Cronin

B.S., Cook College, Rutgers University, 1984

Presented in partial fulfillment of the
requirements for the degree of

Master of Science

UNIVERSITY OF MONTANA

1988

Approved by:

____
Chairman, Board of examiners

____
Dean, Graduate School

____
Date

ABSTRACT

Cronin, Christopher, M.S., June, 1988 Geology

Stratigraphy and Sedimentation of the Ravalli Group (Middle Proterozoic Belt Supergroup) in the Mission, Swan, and Flathead Ranges, Northwest Montana

Director: Don Winston

Correlation of lithofacies within the Burke, Revett, and St. Regis Formations of the Mission Range, eastward through the Swan Range and into the Grinnell Fm. of the Flathead Range demonstrate that the entire Burke-Revett-St. Regis sequence correlates with the Grinnell Fm.

Informal stratigraphic Unit B is a very fine-grained quartzite between the Prichard Fm., and the Burke Fm. and the Burke-correlative interval of the Grinnell Fm. Informal stratigraphic Unit A is a green argillite between the Prichard Fm. and Unit B in the Flathead Range, and is tentatively correlated eastward with the uppermost unit of the Appekunny Fm. in Glacier Park.

The Ravalli Group and Unit B are interpreted to record intracratonic, alluvial apron-sandflat-playa mudflat environments. Episodic floods from a western continent flowed down vast alluvial aprons in braided streams. The flow broadened and shallowed as the apron slope decreased, forming sheetfloods which deposited laterally extensive flat-laminated sand beds, and minor cross-beds. The sheetfloods thinned and spread out across the apron toe, onto sandflats, and finally across vast mudflats, depositing progressively thinner and finer layers of silty-sand and silt. As flood waters ponded and playa lakes spread across the mudflats, clay settled out, forming clays caps. When the playas dried-up, the clay layers were suncracked. Sheetfloods that entered rare perennial lakes on the distal sandflats formed turbid plumes which deposited graded mud layers. Occasional sheetfloods from the east flowed across localized alluvial aprons of coarse sand and spread out across the mudflats and sandflats, depositing laterally extensive beds of coarse sand.

ACKNOWLEDGEMENTS

I am very grateful to Don Winston for introducing me to this challenging topic, and for his energetic support in both the field and the office. Don's tremendous enthusiasm for Belt research inspired me through the course of this project, and will for many years to come.

Thanks to Johnnie Moore and Myer Chessin for serving on my thesis committee. Discussions and correspondencies with Jim Whipple, Jack Harrison, and Jeff Mauk helped my research. A running dialogue with Jeff Kuhn about the relationship between my sections and his Grinnell Fm. sections, and about the depositional environment of the Ravalli Group was very stimulating.

Special thanks to Mike Wolf, Marvin Woods, and Don Winston for their invaluable help in measuring section in very difficult terrain. I have many fond memories of climbing mountains with these friends.

Further thanks to Mike Wolf for photographic work which greatly improved the illustrations.

Thanks to the Belt Association for the research grant which supported my first summer of field work. Additional support was obtained from National Science Foundation grants PRM-801149 and EAR-08409507.

Finally, heartfelt thanks to my wife, Diane, whose love has been my foundation since we first met.

TABLE OF CONTENTS

LIST OF FIGURES

Chapter 1

INTRODUCTION

PURPOSE OF STUDY

Dominantly fine-grained rocks of the Middle Proterozoic Belt Supergroup outcrop across western Montana, Idaho, and eastern Washington, and extend into British Columbia and Alberta, where they are known as the Purcell Supergroup (figure 1). In the United States, the Belt comprises four major stratigraphic subdivisions: 1) Lower Belt, 2) Ravalli Group, 3) Middle Belt Carbonate, 4) Missoula Group (Harrison, 1972). One of the major stratigraphic problems remaining in Belt research is correlating the Ravalli Group from the western to the eastern margin of the Belt basin.

Early workers described stratigraphic sections from three widely seperated areas: 1) the Helena Embayment (Walcott, 1899), 2) Glacier National Park (Willis, 1902), and 3) the Coeur d'Alene District (Ransome, 1905; Ransome and Calkins, 1908) (figure 1). Walcott (1906) correlated a section at Ravalli, Montana with the Burke, Revett, and St. Regis Formations of the Coeur d'Alene District, establishing the group. Correlation has progressed outward from the three centers, but the relationships between the western Ravalli sequence and each of the eastern localities have remained uncertain due to a lack of measured sections or outcrop in the intervening areas

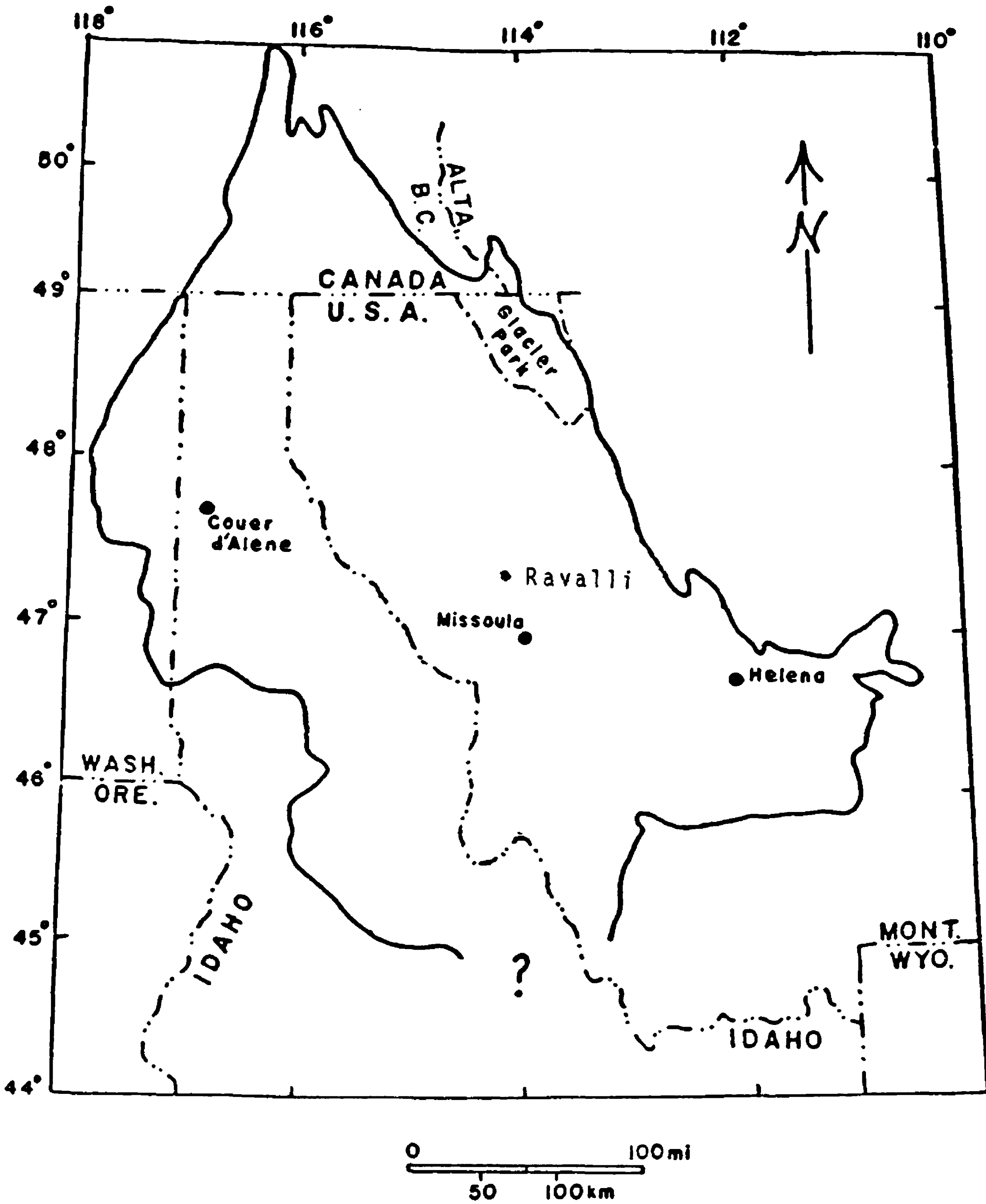


Figure 1. Approximate limit of known Belt outcrop (modified from Harrison, 1972).

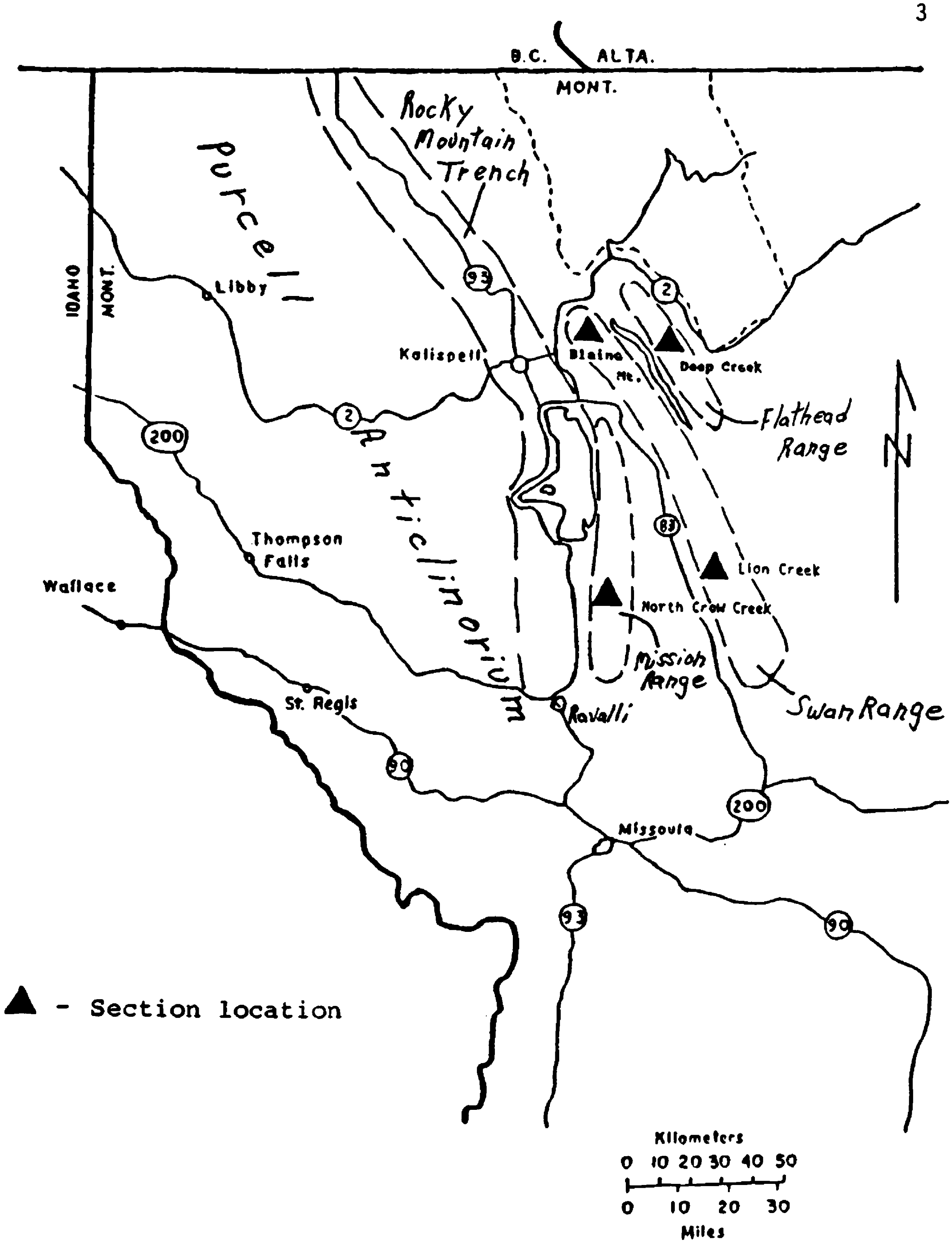


Figure 2. Map showing the locations of the measured sections.

(figure 6). The primary goal of this study is to correlate lithofacies within the Burke, Revett, and St. Regis Formations of the Mission Range eastward through the Swan Range and into the Grinnell (Spokane) Formation of the Flathead Range, thus documenting the stratigraphic relationships between the western and northeastern sequences. Complete sections of the Ravalli Group were measured at North Crow Creek in the Mission Range, Blaine Mountain in the northern Swan Range, and Deep Creek in the Flathead Range. A partial section of the Ravalli Group was measured at Lion Creek in the southern Swan Range (figure 2). The sections were measured with a five foot jacob's staff and sediment-types recorded on a graphical and written log at one inch equals ten feet. Later, the measured sections were redrawn at a scale of one inch equals fifty feet for use in stratigraphic correlation (Appendix E).

The second objective of this project was to interpret the depositional processes of the Ravalli Group sediments in the study area and to synthesize an environmental model in light of the facies trends to the west and northeast. The controversy over whether the Belt was deposited in a marine basin on a trailing continental margin (Price, 1964; Harrison, 1972) or in a landlocked intracratonic basin (Winston et al., 1984; Winston, 1986c; Grotzinger, 1986) has continued from the earliest days of Belt research. While a fluvial-alluvial interpretation of the Ravalli Group's basin-margin quartzites is widely accepted (Alleman, 1983; Mauk, 1983; Greene, 1984; White and Winston, 1982; Bowden, 1977; Horodyski, 1983; Kuhn, 1986; Collins and Smith, 1977), basinward quartzites, siltites and argillites have been interpreted as either tidal flat (Whipple, 1980; Raup et al., 1983) or playa-lacustrine deposits (Winston, 1986c).

GEOLOGIC SETTING

AGE

Radiometric dates of 1430 million years from the Prichard Formation and 1100 million years from the lower part of the Missoula Group are the only constraints on the age of the Ravalli Group (Obradovich et al., 1984).

LITHOLOGY

Rocks of the Ravalli Group are commonly described as agrillite, siltite, and quartzite because diagenesis and greenschist facies metamorphism have altered and indurated the originally muddy, silty, and sandy sediments. In a recent petrologic study of Revett quartzites, Herndon (1983) found "an early authigenic mineral assemblage of quartz-hematite-sericite which indicates Revett sands were initially in an oxidizing and approximately neutral diagenetic environment. Subsequently formed albite-microcline-phengite and later carbonate-chlorite indicate diagenetic waters became progressively more reducing and alkaline as burial proceeded." Widely occurring magnetite, biotite, and rutile were formed during burial and regional metamorphisms (Herndon, 1983). Biotite grade metamorphism in the Prichard Formation passes up to sericite grade metamorphism in the Ravalli Group, which diminishes up-section and toward the east. Sericitic quartzite, siltite, and argillite in the study area possess a weak, spaced cleavage attributed to recrystallization (Herndon, 1983) during Late Cretaceous and Paleocene-Eocene compressional episodes (Sears, 1986). Despite these diagenetic and metamorphic overprints, sedimentary structures are frequently the most striking features of Ravalli Group rocks.

West of the Rocky Mountain Trench, reduced, dark-colored siltites and argillites of the Prichard Formation pass up to a transitional interval of quartzite and siltite characterized by a change from sub-wavebase to shallow-water features (Ransome and Calkins, 1908; Hobbs et al. 1965; Cressman, 1985). The Ravalli Group is dominated by oxidized (grey, purple, red) rocks and characterized by shallow water features (oscillation ripples, sun-cracks). Two major clastic wedges dominated by the fine- to very fine-grained quartzite of the lower and upper Revett seperate three intervals dominated by

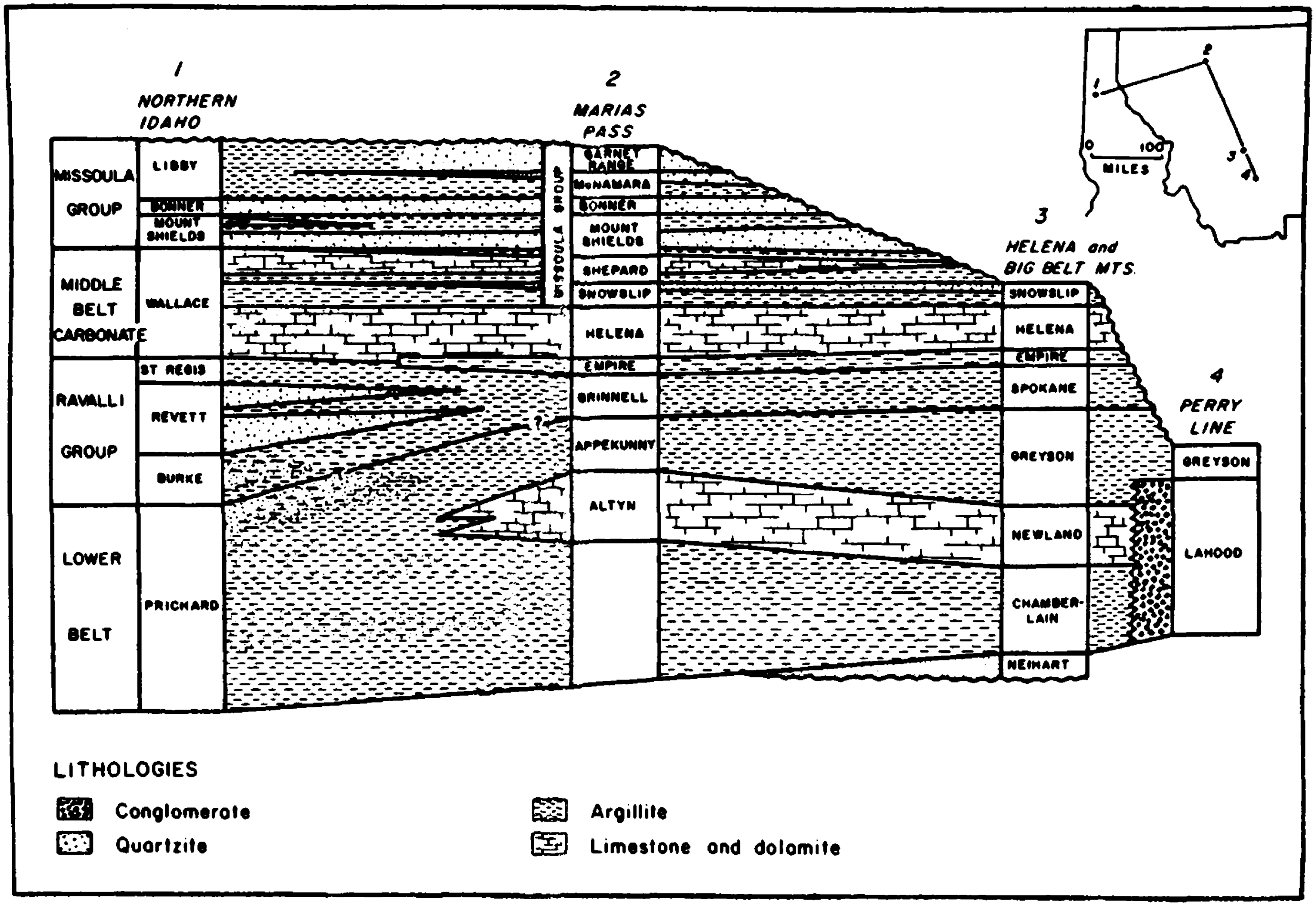


Figure 3. Stratigraphic section showing principal formations of the Belt Supergroup (not to scale; modified from Winston, 1986a).

siltite and argillite belonging to the Burke, middle Revett, and St. Regis (figure 3). The grain-size and bed thickness of all these formations decrease to the northeast, passing to argillite and siltite where they meet interbeds of medium- to coarse-grained quartz arenite, which conversely thin and pinch out westward. This bimodal mixture of mud and coarse quartzite interbeds characterizes parts of the Grinnell and Appekunny Formations in Glacier Park, and the Grinnell (Spokane) Formation of the Swan and Flathead Ranges. Green argillite and calcareous green argillite of the Empire Formation conformably overlie the Ravalli Group across the basin nearly to the Coeur d'Alene District, and records a transtion to the more continuously submerged Middle Belt Carbonate. The Empire has been regarded as a part of the Ravalli Group on the eastern side of the basin, but has recently been tentatively correlated with the lower part of the Wallace Formation to the west (Connor et al, 1984). Because its basinwide correlation is well established, the Empire Formation was not included in this study.

STRUCTURE

Faults active during Belt sedimentation have been identified or postulated (Harrison, 1972; Harrison et al., 1974; Winston, 1986a,d), but no syndepositional faults have been proposed in the study area.

Nearly all Belt strata are allocthonous, having been transported eastward as large thrust slabs. A Late Cretaceous thrust belt trends north-south across the western portion of Montana, while a Paleocene-Eocene, eastern thrust belt forms the Rocky Mountain Front Ranges (Harrison et al., 1980; Winston, 1986a; Sears, 1986) (figure 4). The displacement on the major thrust faults east of the study area is estimated to be 70 kilometers (Earhart et al., 1984). No thrust faults have been mapped between the Mission and Swan Ranges, indicating that these rocks were translated eastward by at least 70 kilometers as part of a large coherent plate moving on a decollment. Thus, the spatial relations between the North Crow Creek, Lion Creek, and Blaine Mountain sections have not be affected by thrust faulting. The structural relationship between these sections and the Deep Creek section in the Flathead range is less certain. While Harrison and others (1980) map a high angle thrust in the Swan and Flathead Valleys, other workers believe that these faults are exclusively of normal displacement (Mudge, 1970; Price and Kluvver, 1974; Winston, 1986a).

Because the Mission, Swan, and Flathead Ranges were formed by normal faulting during the Cenozoic (Winston, 1986a), there is limited structural extension between all the sections in this study. Constenious (1981) attributes 12 kilometers of extension to the Flathead

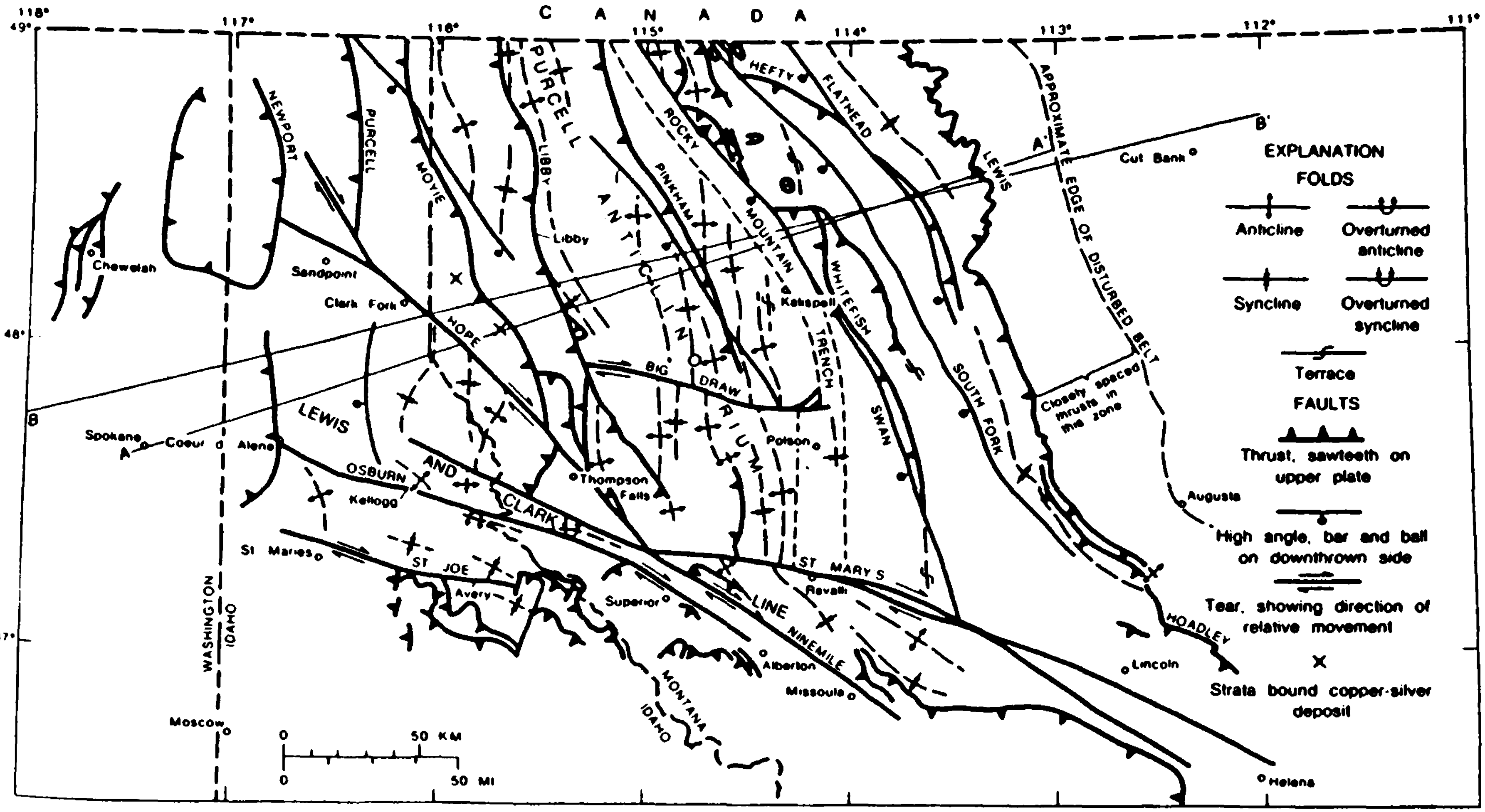


Figure 4. Structure map of the Belt Supergroup in the United States (from Harrison et al., 1980).

Normal Fault between the Whitefish and Clark Ranges, but this is probably an unusually large displacement (Winston, 1986a).

REVIEW OF THE STRATIGRAPHIC PROBLEM

Correlation of the Burke, Revett, and St. Regis formations of the Ravalli Group eastward into the Spokane and Grinnell formations has been a major stratigraphic problem of the Belt Supergroup (Winston, 1986b) (figure 6). According to previous correlations by Harrison (1972), the Burke correlates with the Appekunny Formation, the Revett pinches out in the western part of the basin, and the St. Regis correlates with the Grinnell (Spokane) Formation (figure 5). However, these correlations have remained undocumented because of the lack of measured sections east of the Rocky Mountian Trench.

The Burke and Revett Formations have been mapped from the Coeur d'Alene District, Idaho, across western Montana into the Mission Range (Hobbs et al., 1965; Harrison and Jobin, 1963; Harrison et al., 1986). Mauk (1983) described lower silty and upper quartzitic informal members of the Burke Formation between Idaho and the southern end of the Flathead Indian Reservation. He also included in appendix form a study by Mike Lis describing lower silty, middle quartzite, and upper silty members of the Burke Formation near Kalispell, Montana

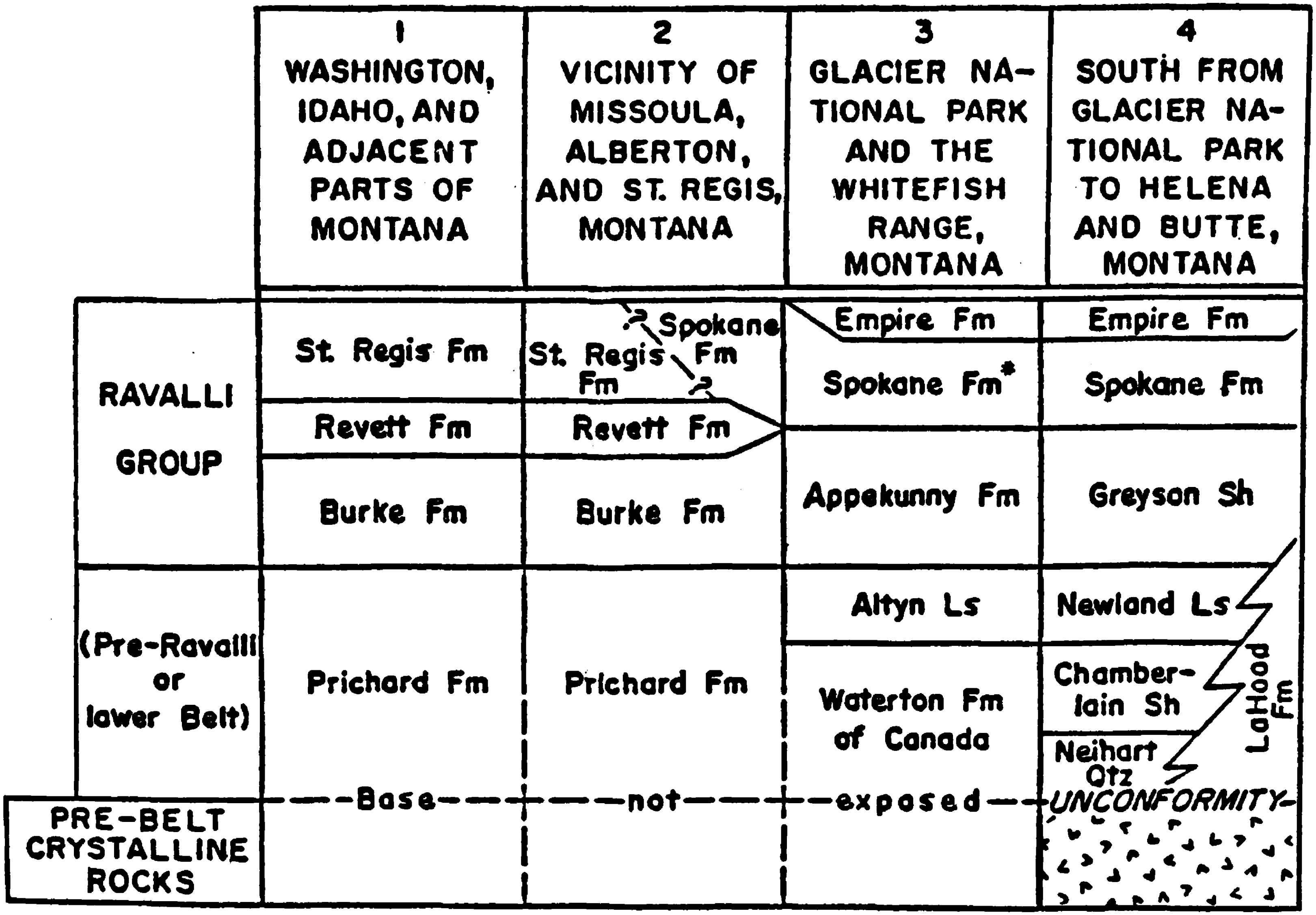


Figure 5. Harrison's correlation of the Ravalli Group and lower Belt (from Harrison, 1972). (* Grinnell Fm. after Whipple et al., 1984; Kuhn, 1986.)

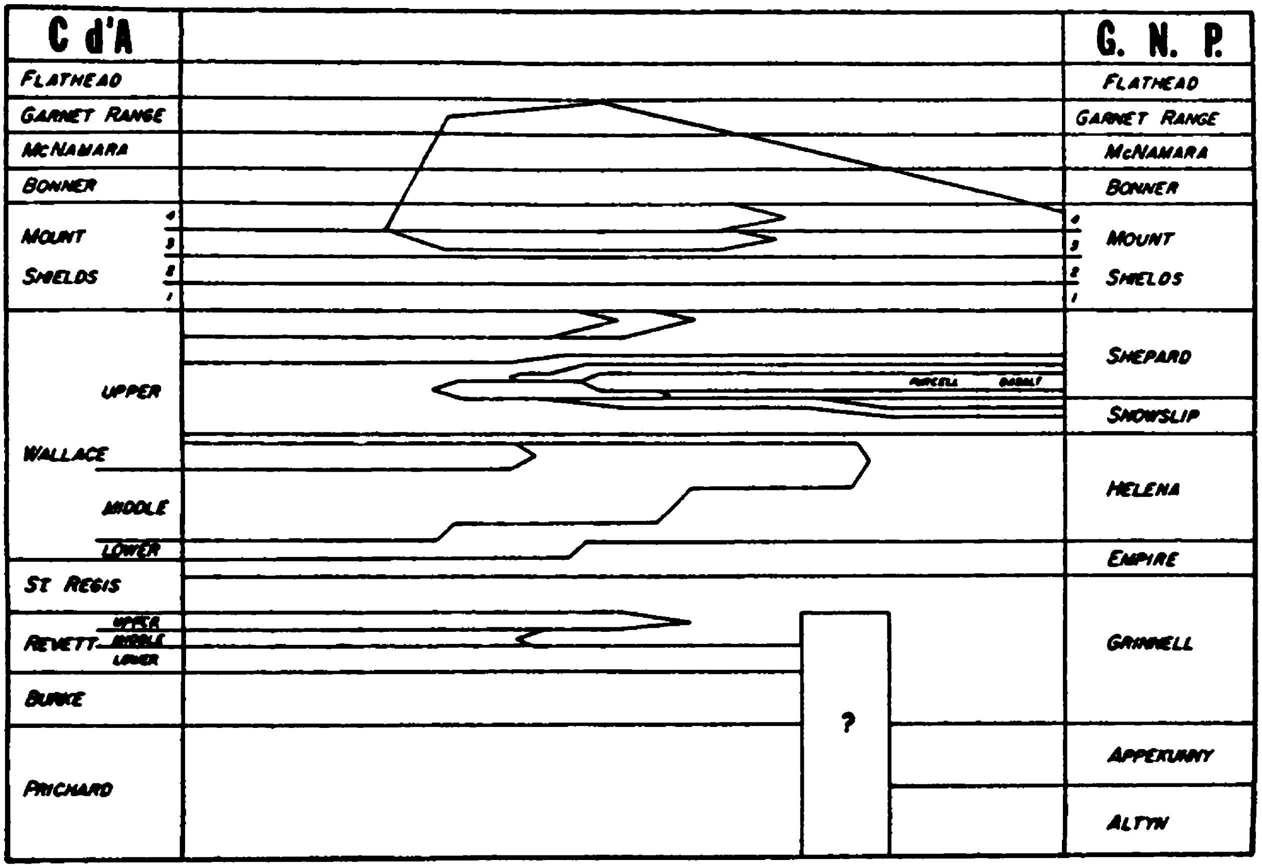


Figure 6. Winston's correlation of the Belt Supergroup between the Coeur d'Alene District (C d'A) and Glacier National Park (G.N.P.); (from Winston, 1986b).

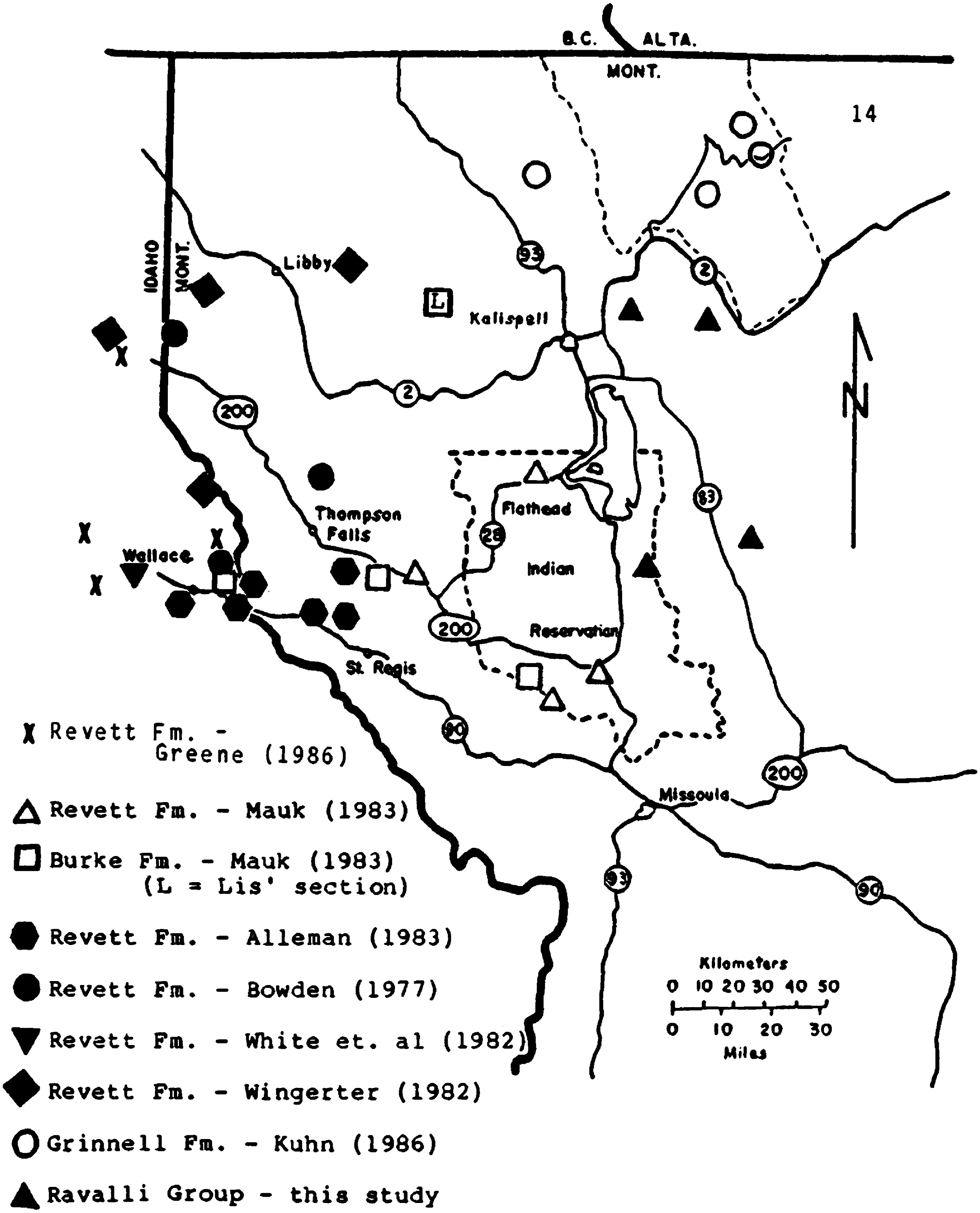


Figure 7. Location of Ravalli Group measured sections in northern Idaho and western Montana (modified from Mauk, 1983).

(figure 7). The relationship between these southern and northern localities of the Burke remains uncertain. White and Winston (1977, 1982) recognized informal lower quartzitic, middle silty, and upper quartzitic informal members of the of the Revett Formation that have been correlated from the Coeur d'Alene District onto the Flathead Indian Reservation (Greene, 1984; Alleman, 1983; Mauk, 1983) (figure 7). West of the Purcell Anticlinorium (figures 2, 4) and east of it on the Montana National Bison Range, silty and argillaceous redbeds above the Revett Formation are mapped as St. Regis Formation. To the east, in the Mission Range, they are mapped as Spokane Formation (Harrison et al., 1983, 1986) (figure 8). The St. Regis and Spokane formations have been reported to be "different sedimentary prisms [derived] from different source directions (Harrison, 1972), although the exact relationship between the two formations (interfingering or overlap) has remained unknown (Harrison 1972; Harrison et al., 1986). Harrison and others (1986) report that "the Spokane Formation is lithologically similar to the St. Regis, except for distinctive white well-rounded medium- to coarse-grained quartzite beds a few inches thick that characterize the unit." However, "where no white quartzite is present, no lithologic criteria are known to distinguish Spokane from St. Regis." (Harrison et al., 1986) and, as a result, the

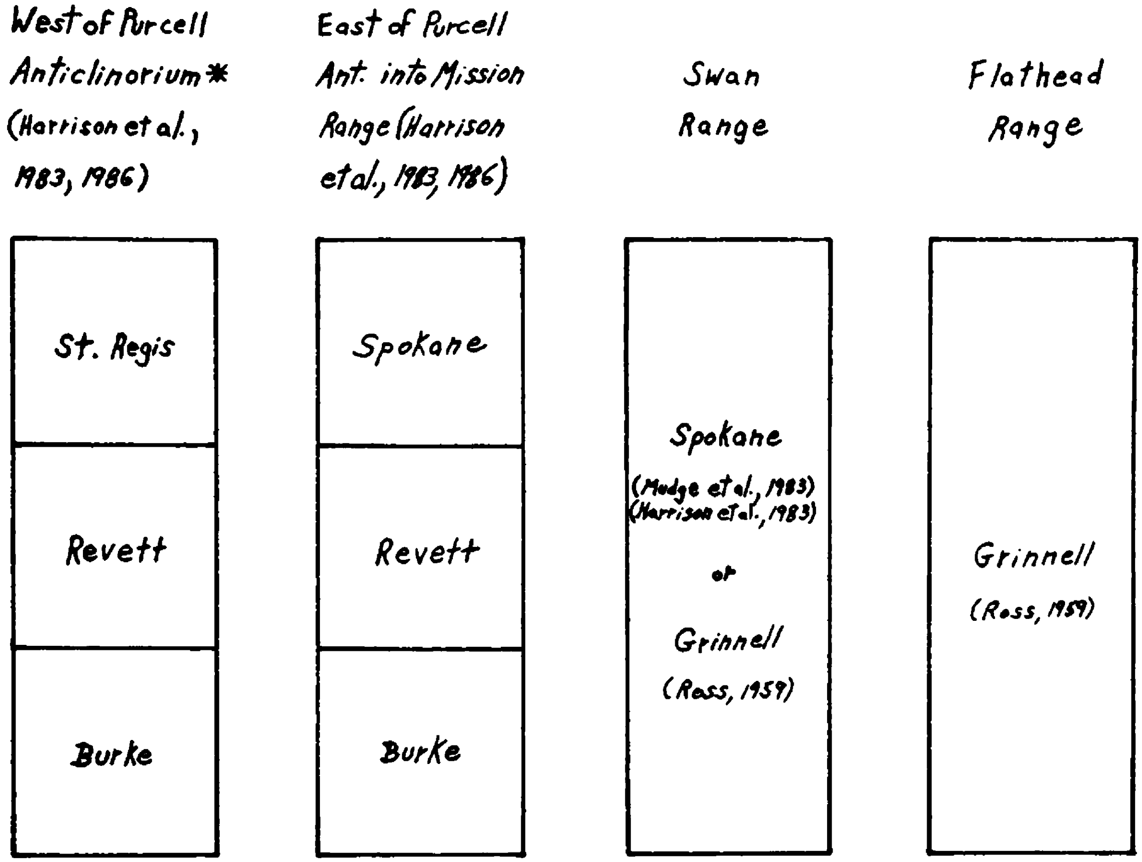


Figure 8. Generalized stratigraphic columns representing the usage of the names St. Regis, Spokane, and Grinnell. * - St. Regis is mapped east of the Purcell Anticlinorium onto the National Bison Range.

name applied to the redbeds above the Revett Formation has often been decided arbitrarily (Harrison et al., 1986). In the Wallace Quadrangle (Harrison et al., 1986) and Kalispell Quadrangle (Harrison et al., 1983), the Purcell anticlinorium is a de-facto arbitrary cut-off (Wheeler and Mallory, 1953; North American Code of Stratigraphic Nomenclature, Article 28c) between the western St. Regis Formation and eastern Spokane Formation.

The names Burke and Revett are not applied east of the Mission Range, and all Ravalli-level strata in the Swan and Flathead Ranges are mapped as the Spokane (Mudge et al., 1983; Harrison et al., 1983; 1986) or Grinnell (Ross, 1959) Formations (figure 8). Thus, the Swan Valley between the Mission and Swan Ranges serves as a second arbitrary cut-off between the western and eastern Ravalli Group nomenclatures. In addition, an unnamed quartzite unit between the Prichard and Spokane Formation near Blaine Mountain in the Swan Range (Harrison et al., 1983) may also belong to the Ravalli Group.

The terminology used for mapping on the east side of the Belt basin is currently in disagreement. The Glacier National Park name Grinnell was placed in synonomy with the Helena Embayment name Spokane by Harrison (1972). The Glacier Park name Appekunny was placed in synonomy with the Helena Embayment name Greyson by Mudge (1977).

However, some workers continue to use the names Appekunny and Grinnell formations (Horodyski, 1983; Whipple et al., 1984; Kuhn, 1986). Since these two pairs of names are applied to the same lithic units, they are essentially interchangable. Because correlations between the Helena Embayment and Glacier Park remain uncertain (Harrison, 1972; Winston, 1986b), and because beds of the coarse-grained quartz arenite characteristic of the park units extend southwestward into the Flathead, Swan, and Mission Ranges, the names Appekunny and Grinnell formations are used in this study.

This study was undertaken in coordination with Jeff Kuhn's (1986) study of the Grinnell Formation in Glacier National Park and the Whitefish Range (figure 7) as a joint-effort to resolve the stratigraphy of the Ravalli Group from Ravalli to Glacier National Park.

Preliminary correlations based on sections in the Mission, Swan, Whitefish, Clark, and Lewis Ranges (Cronin, Kuhn, and Winston, 1986) (Appendix A), supported Winston's (1986b) hypothesis that the entire Burke-Revett-St. Regis sequence correlates with the Grinnell, and perhaps the uppermost Appekunny Formation (figure 6). This study concludes that the Burke-Revett-St. Regis sequence correlates with only the Grinnell Formation (Figure 30).

REVIEW OF DEPOSITIONAL ENVIRONMENT CONTROVERSY

Because Winston (1986c) and Winston, Woods, and Byer (1984) review the controversy over the depositional setting of the Belt, only a summary is included here. Essentially, even the basic tectonic environment of the Belt basin is in question. Some workers consider the Belt to be a miogeocline (Price, 1964; McMechan, 1981) or an embayment of the Proterozoic Ocean (Harrison et al., 1974; Reynolds, 1984), while other workers consider it to be an intracratonic basin (Cressman, 1985, 1988; Winston et al., 1984; Winston, 1986c; Frost and Winston, 1987). Whether the Belt was deposited on a trailing continental margin, or in an inland basin depends on the presence or absence of a continental landmass to the west of the North American craton. A southwestern source is ascribed to the Lower Belt's Prichard Formation (Cressman, 1985), to the Burke- Revett-St. Regis sequence (Harrison, 1972; Mauk, 1983), and to the Middle Belt Carbonate's Wallace Formation (Grotzinger, 1986). In the marine models, the source terrain to the southwest is interpreted as a peninsular or island extension of the North American craton (Harrison et al., 1974), or as the seaward basement rim of a rifted continental margin (Grotzinger, 1986). However, the temporal continuity and volume of the southwestern sediment supply, and the correlation of the Ravalli Group with eastward-thinning quartzite wedges in the Creston Formation of Canada (McMechan, 1981) suggests to some that a continent bounded the entire western side of the Belt basin (Cressman, 1985, 1988; Winston et al., 1984; Winston, 1986c; Frost and Winston, 1987).

One point of unusual agreement in the source terrain controvery is that most workers consider the distinctive, westward-thinning interbeds of medium- to coarse-grained quartz arenite in the Appekunny and Grinnell Formations to have been derived from the North American craton (Smith and Barnes, 1966; McMechan, 1981; Horodyski, 1983; Connor et al., 1984; Kuhn, 1986). However, the source of the argillite and siltite interbedded with the coarser sands is still being debated. While Harrison (1972) proposes that the entire Grinnell (Spokane) Formation is a distinct sedimentary prism derived from the east, Winston (1986c) views the mudstones of the Grinnell as the distal facies of the Burke-Revett-St. Regis sequence, with only the coarse sand coming from the craton.

Sedimentologic interpretations of the Ravalli Group's depositional environment are also sharply divided along the lines of the marine versus intracratonic controversy, and generally propose either a tidal flat or an alluvial/playa-lacustrine model.

In Glacier National Park, Horodyski (1983) and Kuhn (1986) concluded that the Grinnell Formation records deposition on an alluvial plain. However, the Grinnell has also been interpreted as a tidal flat deposit (Raup et al., 1983), as has the Spokane-Empire transition zone of west-central Montana (Whipple, 1980).

The St. Regis Formation was originally interpreted as a turbidite complex (Hrabar, 1973), and the Revett Formation has been interpreted as a barrier island complex (Wingerter, 1982), but most workers (White and Winston, 1982; Bowden, 1977; Greene 1986) conclude that the Revett Formation in Idaho and western Montana represents a braided stream system and associated flood-plain deposits. Working eastward onto the Flathead Reservation (figure 7), Alleman (1983) and Mauk (1983) also interpreted the Revett Formation as a fluvial deposit, but explained copious, laterally extensive flat-laminated sand beds as ephemeral sheetflood deposits. Mauk (1983) concluded that the Revett fluvial system fed a large tide-dominated delta, the tidal mouth bars and upper prodelta of which are represented by the Burke Formation. In contrast, Winston (1986c) proposed an integrated basinwide model of Ravalli Group deposition that involves an alluvial apron-sandflats complex represented by the Revett Formation passing eastward to playa-lacustrine mudflats represented by the middle interval of the Grinnell Formation. He suggests that coarse sand from the craton was transported into the basin by longshore drift along the short-lived shorelines of the repeatedly contracting and expanding lake.

This study concludes that during the deposition of the Ravalli Group, ephemeral braided streams from the south and west passed downslope to sheetfloods that spread out northward and eastward across terrestrial sandflats and then mudflats. Shallow standing waters ponded on the flats after each flood quickly dried up, exposing a vast plain. Occasional sheetfloods from the craton carried coarse sand into the basin as they flowed westward across the mudflats, and even out onto the sandflats on the other side of the plain (figure 32).

Chapter 2

SEDIMENT-TYPES

METHODOLOGY

The stratigraphic sections measured for this study are described in terms of eight sediment-types: 1) cross-bedded sand, 2) flat-laminated sand, 3) even couple, 4) couplet, (with two varieties: lenticular and even couplets, and a sub-variety of even couplets, alpha couplets), 5) mudchip conglomerate, (with sub-types 1, 2, and 3), 6) coarse sand, (with bedded and cross-laminated sub-types), 7) bedded silt and sand, and 8) microlamina. Each sediment-type represents a recognizable sediment taxon based on sedimentary structures, grain-size, and inferred sedimentary mineral composition (Winston, 1986c). The sediment-type methodology was introduced by Winston (1986c) as a means for describing and interpreting Belt strata on a sedimentologic basis, rather than on the mixture of sedimentology, diagenesis, and metamorphism which forms the basis of formal units in the Belt. This new technique was first used to interpret the sedimentation of the Ravalli Group, Middle Belt Carbonate, and the Missoula Group (Winston, 1986c), and has subsequently proven effective in stratigraphic/sedimentologic studies of the Shepard (Woods, 1986) and Grinnell Formations (Kuhn, 1986).

The merits of using sediment-type terminology rather than the rock-type vernacular used in mapping and other geologic studies (Winston, 1978, for example) are reviewed by Winston (1986c) and Woods (1986). Basically, formations are defined on the basis of grain-size, composition, and color. Lithic descriptions include these features, plus whatever descriptive adjectives are deemed appropriate; for example: white cross-bedded quartzite, wavy-bedded purple siltite, green laminated argillite. Whereas lithology is a proper basis for mapping, lithologic nomenclature does not convey enough information for detailed sedimentologic correlations and facies interpretations. Winston (1986b) and Lemoine and Winston (1986) have shown that formal stratigraphic units cut across sedimentologically correlated units.

By emphasizing sedimentary features instead of total lithology, sediment-types focus attention on the primary, depositonal relationships between bodies of sedimentary rock. Since most field descriptions include color, in addition to sediment-type, they can be readily translated into formal stratigraphy, whereas the reverse procedure is very difficult, and fraught with uncertainties, because formal stratigraphic units are not based on sedimentary structures.

A brief overview of the Ravalli Group's sand component, and the nature of the sediment-types follows. Then, each sediment-type is described and interpreted in terms of its own depositional processes. The sequence of processes that deposited each sediment-type are analyzed, and then considered in light of the hydrodynamic conditions inferred for the adjoining sediment-types. The sediment-type environmental interpretations are synthesized into a depositional model in chapter 4 of this thesis.

OVERVIEW

Two distinct, mutually exclusive sand populations occur in the Ravalli Group. In both outcrop and thin-section, the medium- to coarse-grained, rounded to very well rounded, slightly feldspathic (5 to 10% feldspar) sand of the coarse sand sediment-type contrasts sharply with the very fine- to fine-grained, sub-angular to sub-rounded feldspathic (10 to 20% feldspar) sand of the finer sand population that characterizes the cross-bedded sand, flat-laminated sand, even couple, and bedded silt and sand sediment-types, and occurs in the silty part of couplets. Figure 9 contrasts the size and shape of the coarser and finer sands in a rare mixture of the two at a bed contact.

The flat-laminated sand, even couple, and couplet sediment-types subdivide a continuum of graded beds into thickness-ranges of 150-10 (flat-laminated sand), 10-3 (even couples), and 3-0.5 centimeters (couplets). The boundaries between these three orders of graded bedding

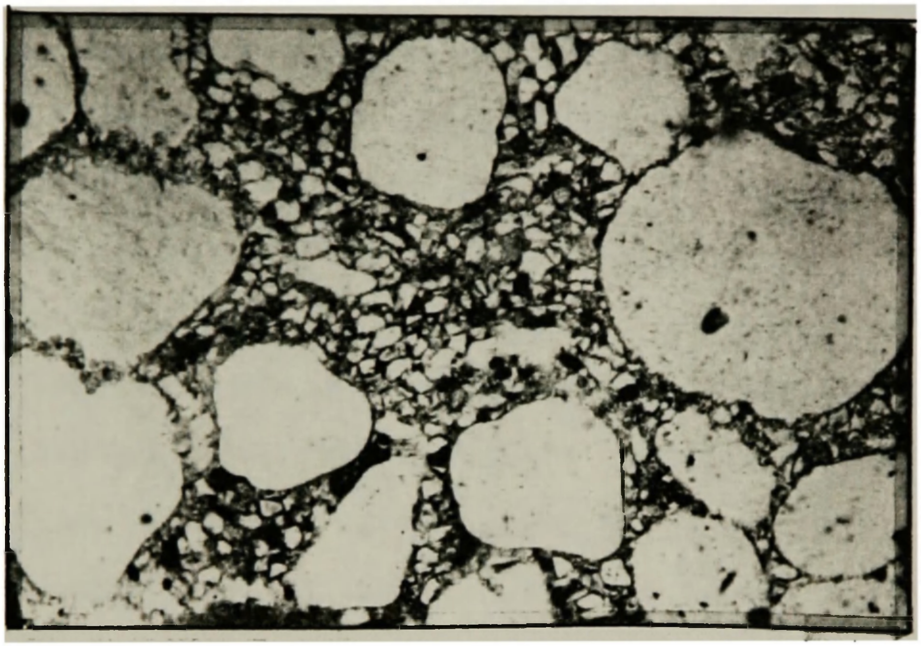


Figure 9. The two sand species -- Photomicrograph showing a mixture of sub-angular to sub-rounded, very fine-grained sand and rounded to very well-rounded, medium- to coarse-grained sand. Thin section of the interface between a flat-laminated sand bed and a bed of the coarse sand sediment-type.

are useful because beds in these size-ranges tend to occur together in sets, and the increase in the proportional thickness of mud layers with decreasing bed thickness causes bed appearance to change at about these thickness boundaries. For example, flat-laminated sand beds commonly have a very thin mud cap, or none at all; in the even couple size-range, mud layers normally occur and constitute a significant portion of the beds; on the scale of couplets, mud generally dominates over silty layers. Differential weathering and cleavage in the Ravalli Group cause the muddy layers capping beds of these sediment-types to recess, allowing strata thickness to be determined even from the lichen-encrusted outcrops in some parts of measured sections. In many exposures, the weak, spaced cleavage refracts through the muddy layers, dying out in the silty or sandy layers. In contrast to the graded bedding continuum, the cross-bedded sand, mudchip-conglomerate, bedded silt and sand, microlamina, and coarse sand sediment-types are each distinct taxon.

CROSS-BEDDED SAND SEDIMENT-TYPE

DESCRIPTION

This sediment-type includes all large-scale cross-bed sets composed of foreset cross-laminae of the very fine- to fine-grained sand population. Two distinct set geometries are visible in the dominantly two-dimensional outcrops: sets bounded by essentially flat, parallel surfaces are referred to as planar-tabular sets, while those in which cross-laminae toes meet scoop-shaped basal contacts are identified as trough sets (terminology of Allen, 1982, vol.1, p. 348) (figure 10).

Planar-tabular sets range from 15 to 70 centimeters-thick and can often be traced across 10 or more meters of outcrop before exposure is covered, cross-laminae can no longer be discerned, or the set gradually thins and passes to flat-laminated sand. Cross-laminae of any single planar-tabular set are stikingly uniform, but commonly vary from set to set. Cross-laminae are from 1 to 3 centimeters thick, and range from straight with angular toes (figure 11), to straight with tangential toes, to concave with sweeping lower contacts (figure 12 and figures 10a,b,c). Straight-angular cross-laminae have non-erosional basal contacts, but whether straight-tangential and concave cross-laminae meet a scoured or depositional planar surface is indeterminate. The dip angles of cross-laminae in planar-tabular sets range from 13 to 44

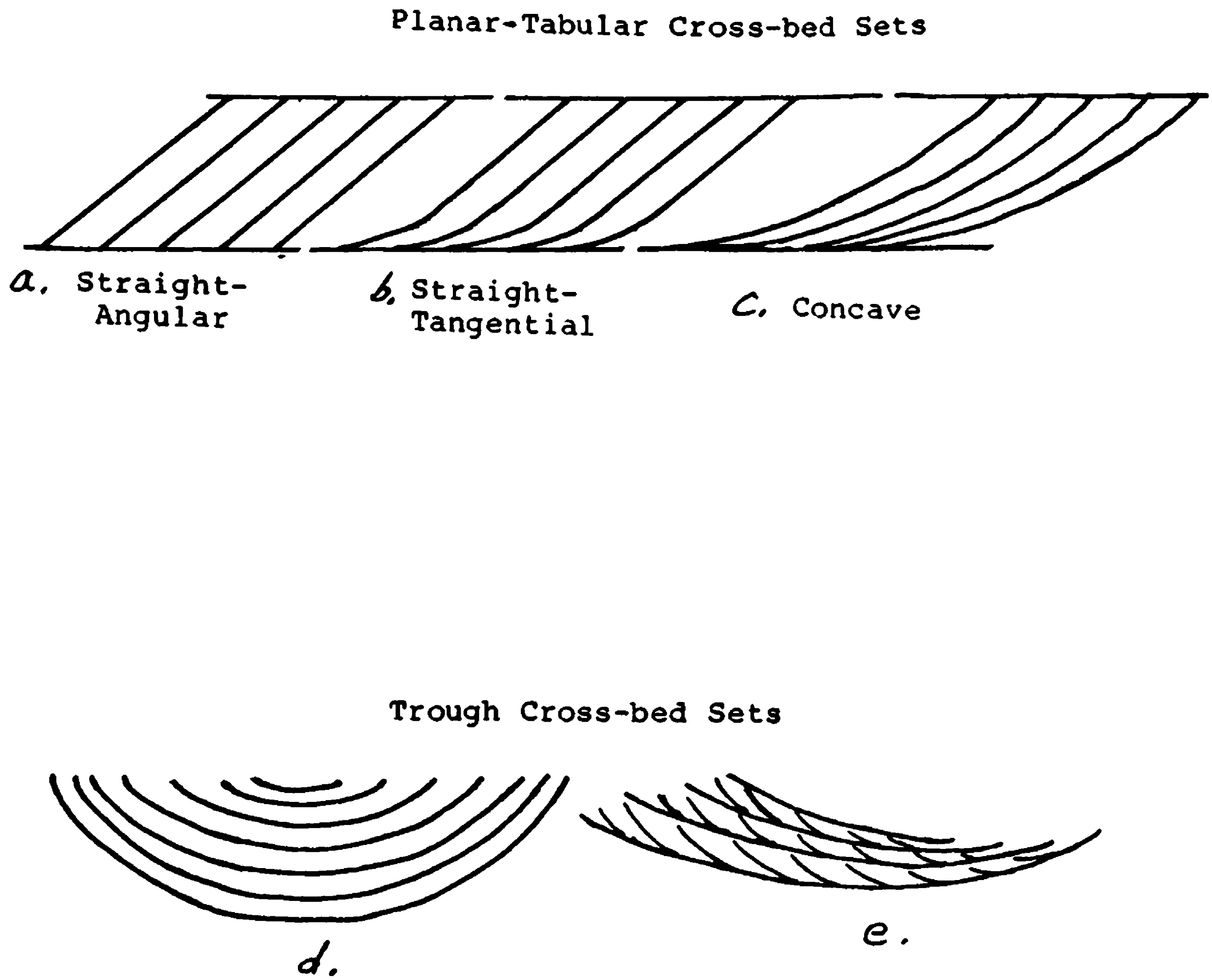


Figure 10. Geometry of planar-tabular and trough cross-bed sets: a,b,c - planar-tabular cross-bed sets showing variations in the nature of cross-laminae; d - trough cross-bed set shown perpendicular to flow direction; e - three stacked trough cross-bed sets (a coset) shown parallel to the flow direction.

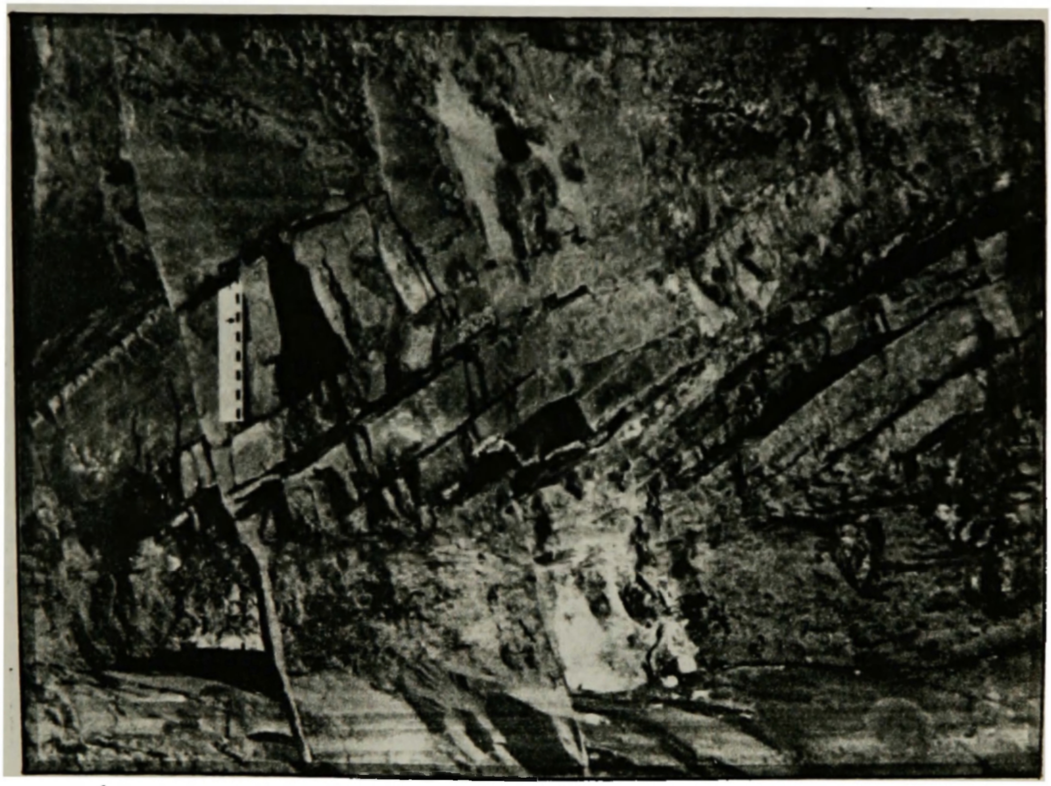


Figure 11. Cross-bedded sand sediment-type.
Planar-tabular set with straight-angular cross-laminae.

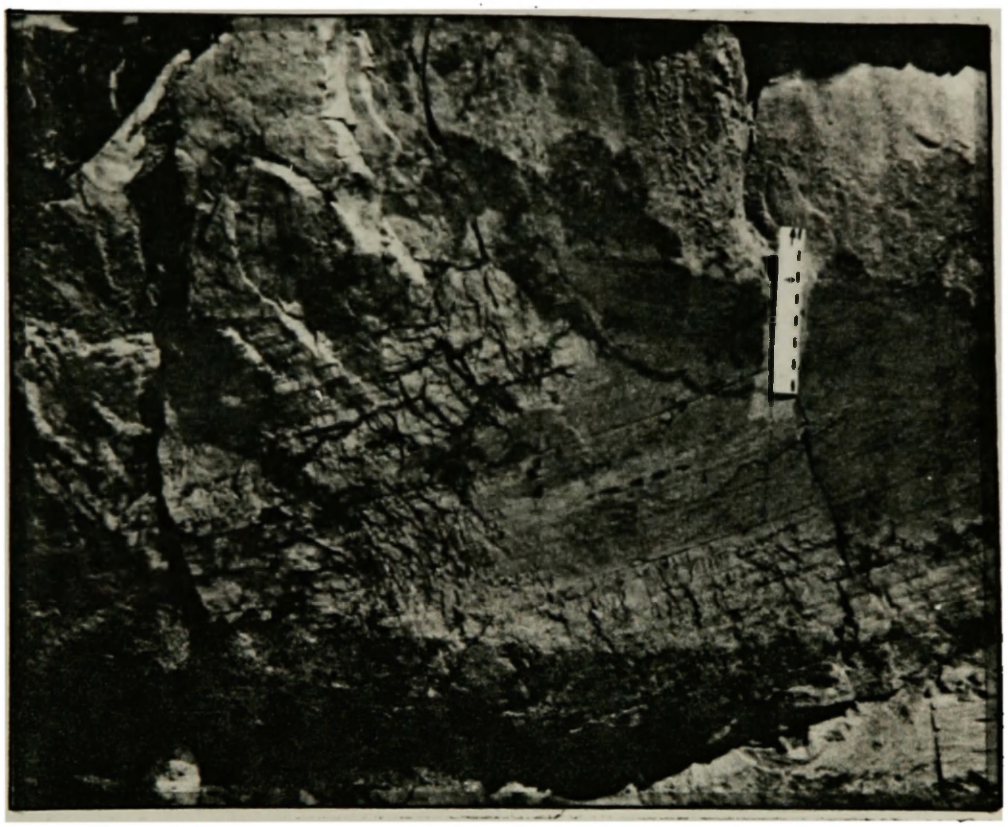


Figure 12. Cross-bedded sand sediment-type. Planar-tabular set with sweeping concave cross-beds. Set thins to the right. Note topset bed of flat-laminated sand (perpendicular to ruler).

degrees and average 28 degrees, based on forty-nine attitude measurements corrected for tectonic tilt (table 1). In most planar-tabular cross-bed sets the cross-laminae are more clearly defined than their upper parts and the cross-laminae are vaguely reverse-graded. However, in some cross-bed sets, cross-laminae are distinct from the upper to the lower set contact, while in others the cross-laminae are indistinct along their entire incline.

Only six trough cross-bed sets were observed in the measured sections. The three trough sets in the lower Revett Formation at North Crow Creek are exposed perpendicular to the flow direction (figure 10d). The sets range from 1.0 to 2.5 meters wide and from 0.3 to 0.6 meters thick. The three stacked trough cross-bed sets (figure 10e) in the upper Revett correlative interval of the Grinnell Formation at Lion Creek (Appendix E) are exposed parallel to the flow direction. The lowest set is the largest and the upper sets are progressively smaller, as illustrated in figure 10e.

No reactivation surfaces were observed in any of the cross-bed sets. Some cross-bed sets pass up into flat-laminated sand without a reactivation surface (figure 12).

Sets of the cross-bedded sand sediment-type are interbedded with the thicker-bedded parts of flat-laminated sand intervals and they have two modes of occurrence: 1) solitary cross-bed sets, and 2) cosets composed of stacked cross-bed sets. The three trough cross-bed sets in the North Crow Creek section are solitary, while the three in the Lion Creek section form a coset. Most of the planar-tabular sets at North Crow Creek are solitary. The mode of occurrence and foreset cross-laminae shape of planar-tabular sets in the Revett Formation at North Crow Creek are shown in table 1.

The mode of occurrence and cross-bed shape of planar-tabular sets in the Revett Formation at North Crow Creek are shown in table 1. No reactivation surfaces occurred in any of the planar-tabular or trough cross-bed sets observed in this study.

INTERPRETATION

Solitary sets of the cross-bedded sand sediment-type are interpreted to be the product of isolated straight- or sinuous-crested large-scale bedforms, whereas, cosets were probably deposited by trains of accreting large scale bedforms. The uniformity and continuity of cross-laminae within sets of the cross-bedded sand sediment-type indicate that deposition occurred by two interactive processes on the slipfaces of large-scale bedforms. They are grainfall and grainflow. Grainfall is "the settling out of grains through a fluid moving too slowly to form ripples on the depositional surface", while grainflow is "a form of sediment gravity flow in which the grains are supported by collisions between grains" (Hunter and Kocureck, 1986). Much of the bed and suspended load carried from the stoss sides of large-scale bedforms onto their lee sides settles onto the slipfaces by grainfall. Grainfall deposits build up until the angle of initial yield is reached and grains flow down the slipface, remixing the grainfall sediments into grainflow laminae (Reineck and Singh, 1980; Hunter and Kocureck, 1986).

The indistinct upper portions of most cross-laminae in sets of the cross-bedded sand sediment-type indicate that grainflow high on the slipface was nearly continuous because of the heavy grainfall sedimentation there, while the more distinct lamination lower along cross-laminae reveals deposition by intermittent grainflows which produced reverse-graded cross-laminae as a result of shear-sorting during grainflow (Reineck and Singh, 1980; Hunter and Kocureck, 1986). Cross-laminae distinct along their entire incline are either erosional remnants, or the product of a relatively lower sedimentation rate and flow velocity, as indicated by the intermittent grainflow sedimentation. Where cross-strata are indistinct along their entire incline, a relatively higher sedimentation rate and flow velocity are confirmed by the continual nature of grainflow and grainfall sedimentation (Reineck and Singh, 1980; Hunter and Kocureck, 1986).

The shape of cross-laminae indicates the nature of flow at the front of the bedforms and thus suggests their morphology. Straight foresets with angular contacts indicate that grainflows halted when they attained a stable angle of repose and that the lee-side eddy did not redistribute the grains. Straight cross-laminae with tangential toes and concave cross-laminae reflect a stronger lee eddy that swept sediment back onto the base of the slipface (Jopling, 1965; Reineck and Singh, 1980). This suggests that sets with straight foresets were generally deposited in relatively slower and/or deeper flows than those with more concave foresets. Since increasing flow velocity causes bedforms to develop greater sinuosity (Harms et. al, 1982), and lee-eddy vigor increases with greater obliquety of the slipface to the flow direction (Hunter and Kocureck, 1986), sets with concave foresets were probably deposited by bedforms with curved or sinuous fronts. The trough cross-bed sets represent the continuation of this progression in which a curved scour pit developed in front of the bedforms. Given that sets of the cross-bedded sand sediment-type range from tabular with straight to concave cross-strata, to well-developed troughs, it is clear that a complete spectrum from straight-crested to sinuous-crested bedforms deposited the cross-bedded sand sediment-type.

These large-scale cross-bed sets could represent regularly spaced megaripples formed under lower flow regime equilibrium conditions (Simons et. al, 1965), or could represent solitary or irregularly arranged bars unrelated to the flow regime (Smith, 1972; Miall 1977, 1985; Blodgett and Stantley, 1980). Because the migration of an isolated bar normally produces just a single set of cross-beds, solitary tabular and trough sets are intepreted to be the deposits of isolated, straight- to curve-fronted bars unrelated to the flow regime. This conclusion is supported by the occurrence of a topset bed of flat-laminated sand on many solitary cross-bed sets. Such topsets are a normal part of bars growing downstream (Crowley, 1983), but are not characteristic of megaripples.

Cross-bed cosets could be produced by a group of bars in which topset beds were scoured during migration, or they could be produced by a train of megaripples. The coordinated downstream shrinkage of the three sets in the Lion Creek coset described above (figure 10e) suggests that these three bedforms were megaripples hydrodynamically related to each other during waining and shallowing flow (Jones, 1977; Rubin and Hunter, 1982). However, such coordinated behavior is not apparent in other cosets, and deposition by groups of bars is favored.

In contrast to the flow regime requirements of megaripples, bars can develop wherever lateral or vertical flow expansion occurs (Cant and Walker, 1978), because rapid deposition on the lee faces creates topography, which in turn increases flow seperation, thus building up a large slipfaced bedform (Hunter and Kocureck, 1986). In this manner, a bed irregularity, or simply the spreading-out of a flow can give rise to a bar.

The absence of reactivation surfaces in sets of the cross-bedded sand sediment-type suggests that they were deposited in individual flood events characterized by flow deceleration without rapid fluctuations or flow reversals. Although the downstream termination of only a few bedforms was observed, the gradual thinning of these sets and their passage to flat-laminated sand suggests that the bedforms simply flattened as the flow seperation decreased with the waining and shallowing of the flow. In other words, the bedforms were not left stranded or reactivated. Rather, they developed, advanced, and flattened within single flood events. The 15-70 centimeter-thick cross-bed sets indicate that flow was at least several deciemeters to one meter deep.

Even where sets and cosets of the cross-bedded sand sediment-type are most abundant, they are still a minor constituent of the section; the dominance of flat-laminated sand beds suggests that sheetfloods were the dominant mode of fluid flow (see the flat-laminated sand sediment-type). Stratigraphic relations imply that braided rivers flowing down vast alluvial aprons flowing from the south, southwest, and west into the study area (Bowden, 1977; Alleman, 1983; Mauk, 1983; Greene, 1984; Winston, 1986c), spread from distinct channels, to the point that they are better described as streams of braided flow within proximal sheetfloods, analagous to floods reported by McKee and others (1965) and Stear (1985). As the floods spread out across the toes of the alluvial apron and out onto vast sandflats represented by the flat-laminated sand sediment-type, they decellerated and deposited isolated bars or groups of bars.

The following paleocurrent analysis further assesses the nature of the flow system, and indicates that the direction of net sediment transport was to the northeast.

PALEOCURRENTS

Thirty tabular sets of the cross-bedded sand sediment-type in the lower Revett and two in the upper Revett of North Crow Creek yielded forty-nine foreset attitude measurements. This was the only section which contained enough cross-beds to warrant a systematic investigation. A detailed description of the paleocurrent analysis is included in Appendix B:

Paleocurrent Methodology and Data. Basically, the data were corrected for tectonic tilt as per Potter and Pettijohn (1977), while the vector mean azimuth (0) and the consistency ratio (L) for each set were calculated according to the widely-used method of Curray (1956). The vector mean, 0, is a measure of the central tendency of the input azimuths, while the consistency ratio, L (0-100%), is a measure of the dispersion of measurements about the vector mean. When L = 100%, all the data are the same; when L = 0%, the data is uniformly distributed about the vector mean. Table 1 includes 0 and L for each set.

A frequency distribution divided into twenty degree intervals (0-19, 20-39, etc.) employed the vector mean azimuth of each set. Because solitary sets and cosets may record hydrodynamically different bedforms, paleocurrent roses were constructed for each type seperately, and also combined. The diagrams are constructed with the line segment radius directly proportional to the square-root of the interval percentage, so that the value plotted is directly proportional to the interval area, rather than to the segment length (Miall, 1974; Greene, 1984). Paleocurrent diagrams for a) solitary sets, b) cosets, and c) solitary sets + cosets are in figure 13.

A fundamental assumption of this analysis is that

Footage	Corrected Attitude	0	L	Set Thickness	Foresets	Occurrence
6468	020/21se 006/21se 064/29se 046/32se	124	92	45 cm	C	S
6412	330/28ne	060	100	45	ST	S
5292	320/42ne 306/38ne 281/31ne	032	96	40	C	S
5265	074/24nw	344	100	10	ST	S
5010	025/22se	115	100	30	ST	S
4850	330/31ne 286/21ne 079/23nw 327/29ne	031	87	70	C	S
4800	358/34ne	088	100	60	ST	S
4795	008/30se 010/24se	099	100	40	C	S
4792	359/35ne 316/22ne	068	93	30	SA	S
4758	004/39se 348/33ne	086	99	60	ST	S
4650	059/30nw 281/25ne 029/23nw	333	87	70	ST	S
4635	346/24ne	076	100	30	ST	S
?	281/26ne 312/14ne	027	96	30	C	S
?	314/28ne 083/13nw	019	90	20	C	S
?	321/17ne	051	100	30	C	S

Table 1: Paleocurrent data and statistics on tabular cross-bed sets in the Revett Formation at North Crow Creek.

?	333/30ne	063	100	30	ST	
						< D
?	080/22nw	350	100	50	ST	
?	359/25ne	089	100	40	ST	
						< D
?	342/38ne	072	100	30	ST	
						< D
?	354/28ne	084	100	40	ST	
?*	338/31sw	248	100	35	ST	
						< D
?*	328/27sw	238	100	20	SA	
						< D
?*	005/21sw	275	100	25	ST	
						< D
?*	334/14sw	244	100	40	ST	
?*	337/24sw 319/30sw 326/30sw	237	99	20	C	
						< D
?*	355/36sw	265	100	15	C	
						< D
?*	349/31sw	259	100	6	C	
?*	337/39ne	067	100	30	SA	S
?	355/43ne	085	100	70	C	S
?	352/44ne	082	100	25	C	
						< D
?	331/25ne	061	100	15	SA	
?	349/40ne	079	100	20	ST	S

Table 1: Paleocurrent data and statistics on planar-tabular sets of the cross-bedded sand sediment-type in the Revett Fm. at North Crow Creek.

O: vector mean azimuth; L: consistency ratio
Foresets: SA - straight angular; ST - straight tangential; C - concave.
Occurrence: S - solitary set; < D - direct contact between sets forming a coset.
? - are listed by ascending order in the section; all are between 4500 and 5000 ft. in the lower Revett section.
* - all these are located in an interval 4 meters thick.

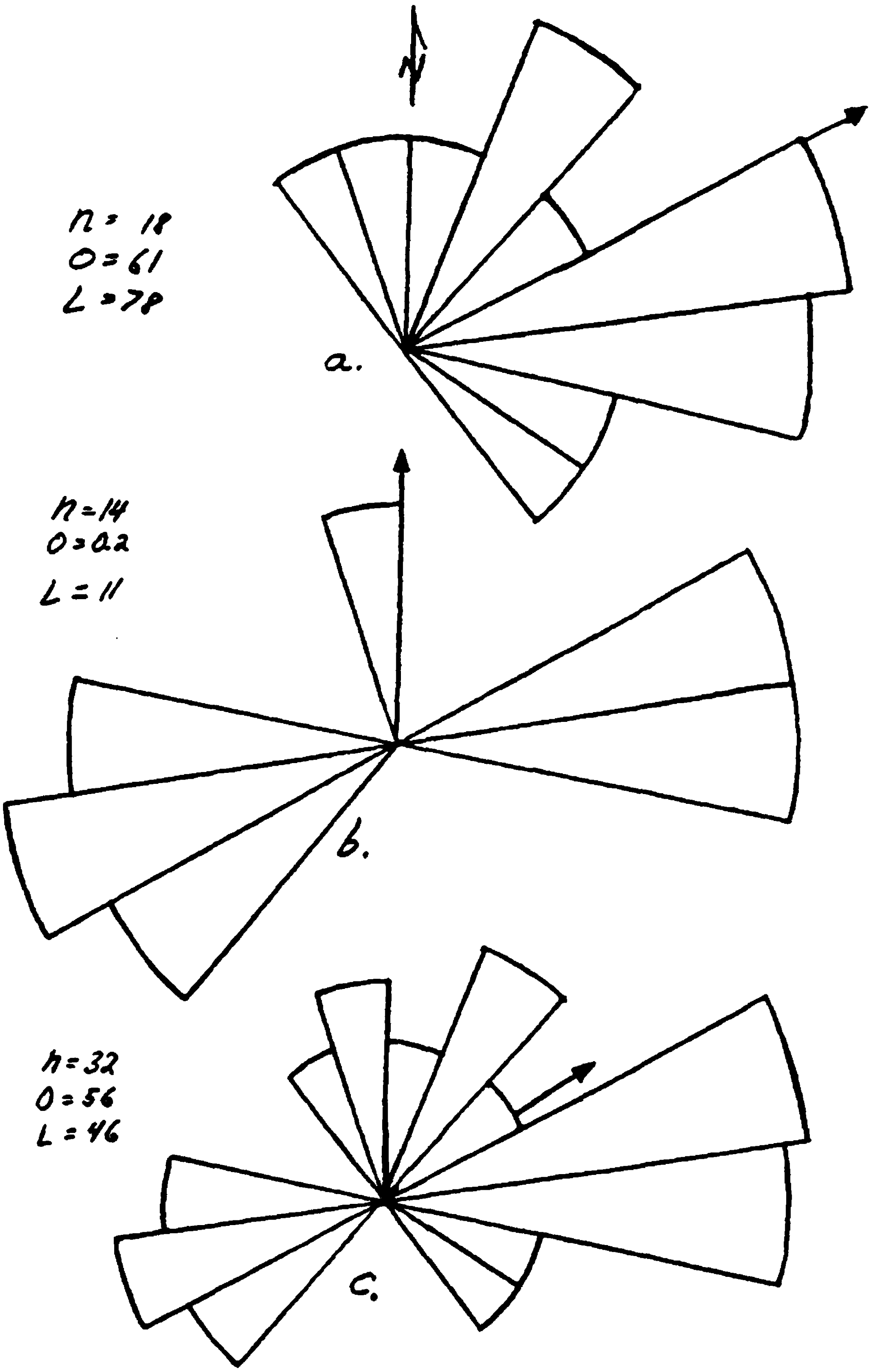


Figure 13. Paleocurrent frequency distribution roses for planar-tabular sets of the cross-bedded sand sediment-type in the Revett Formation at North Crow Creek: a) solitary sets, b) cosets, c) solitary sets + cosets. n = number of sets, O = mean vector azimuth (arrow on rose diagram), L = consistency ratio.

the flow direction is perpendicualr to the strike of the foresets. Studies in ephemeral (Williams, 1971) and braided streams (Smith, 1972) indicate that flow actually crosses a high proportion of bar slipfaces at a large angle, producing a wide scatter of foreset azimuths (Smith, 1972; High and Picard, 1974). For example, Smith found that "only 30.5% of the foresets measured are oriented within 5 degrees of the current which formed them; 42.6% of the dip directions range within 10 degrees, 61.9% within 20 degrees, 72.8% within 30 degrees, and 86.1% within 45 degrees." This suggests that although the tabular set azimuths in this paleocurrent study differ somewhat from the mean flow direction, as a group, they represent the dispersion about the mean flow direction (Smith, 1972). Since the solitary set data in figure 13a were gathered from more than a thousand feet of section, yet does not evidence systematic changes in flow direction with time, their vector mean azimuth (61 degrees) probably reflects a regional paleoslope and records net sediment transport to the northeast.

In the coset rose diagram (figure 13b), the northeast azimuths support the existence of a dominant paleoslope to the northeast, while the northwestern azimuth is consistent with the dispersion of flow on that surface. The southwestern azimuths are derived from two cosets, one with four cross-bed sets and the other with three sets, that occur within a four meter interval (see table 1). They suggest localized flow opposed to the regional paleoslope. Since ephemeral sheetfloods flowing onto flat surfaces spread out by the driving force of their hydraulic head, they may flow in many directions. Because the water surface slope affects the flow direction more than the ground slope (Mader and Teyssen, 1985; Alam et al., 1985), "flow reversal phenomena may be a characteristic of very low gradient fluvial systems" (Alam et al., 1985). Indeed, bipolar cross-beds, and even herringbone cross-beds have been reported from ephemeral fluvial deposits (Mader and Teyssen, 1985; Alam et. al, 1985). They differ from tidal herringbone cross-beds by lacking reactivation surfaces, and the lateral accretion deposits characteristic of tidal channels.

FLAT-LAMINATED SAND SEDIMENT-TYPE

DESCRIPTION

The flat-laminated sand sediment-type in the Ravalli Group is characterized by 10 to 150 cm thick, laterally continuous and even, fining-upward beds composed of flat, parallel laminations of fine- to very fine-grained sand or silty sand and compares closely to Winston's (1986c) definition of the sediment-type. Tabular flat-laminated sand beds can be traced across outcrops more than 30 meters wide, and extend even farther in the walls of the North Crow Creek and Lion Creek canyons. Internal laminations of beds also extend laterally without interruption or reactivation.

While thicker (30-150 cm) beds commonly consist of clean sand from base to top, upward-increasing admixtures of silt, and to a much lesser extent clay, occur in the upper few decimeters of many thicker beds. In the field, the lower parts of some thicker beds appear structureless, but in most, uniform, decimeter- to meter-long flat-laminations formed by subtle grain-size variations (figure 14) are discernable throughout, and in some places are enhanced by heavy mineral laminae.

Most thinner (10-30 cm) flat-laminated sand beds include upward-increasing admixtures of silt and minor clay. The most distal thinner beds contain up to fifty percent silt and perhaps two or three percent clay. Thinner beds fine upward, and flat-laminations are visible throughout most of these beds.

Basal contacts of flat-laminated sand beds are sharply defined above the finer tops of the underlying beds. Most of the basal sand contacts are remarkably flat with little evidence of scour (figure 14). Rare undulating surfaces with a few centimeters of reflief record slight scour. More commonly, symmetrical or asymmetrical ripples are preserved on bedding surfaces,

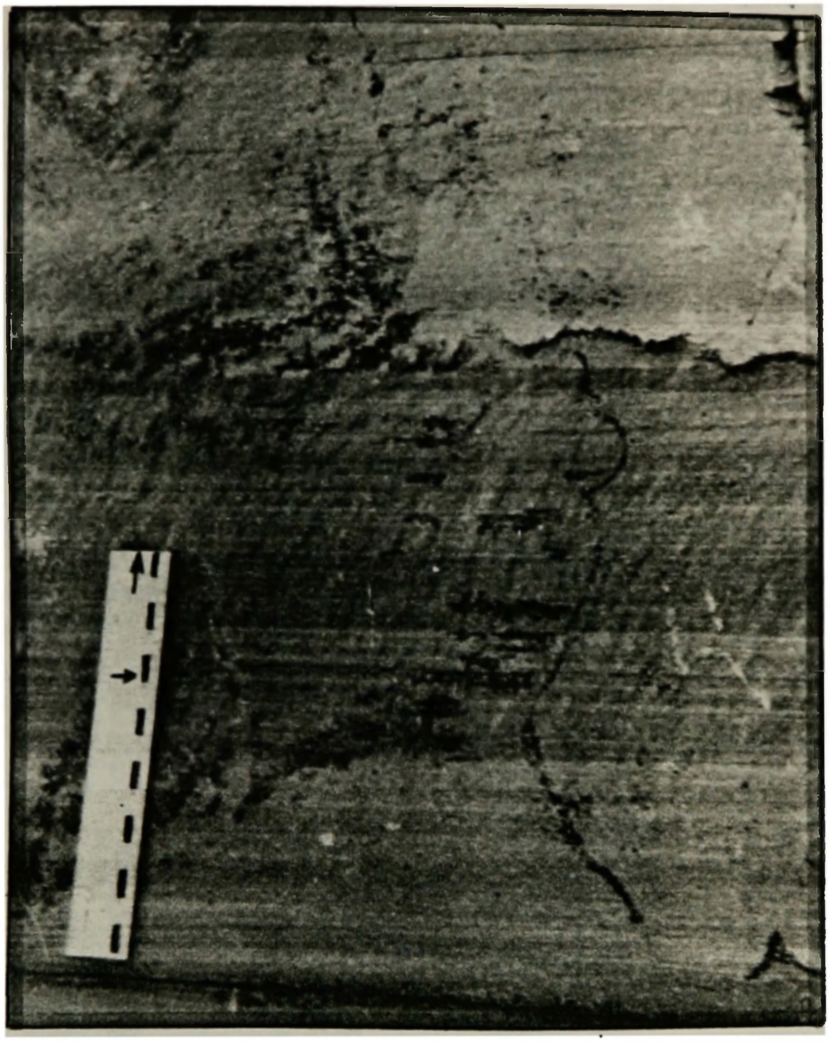


Figure 14. Flat-laminated sand sediment-type. Flat-laminated sand beds with planar contacts.

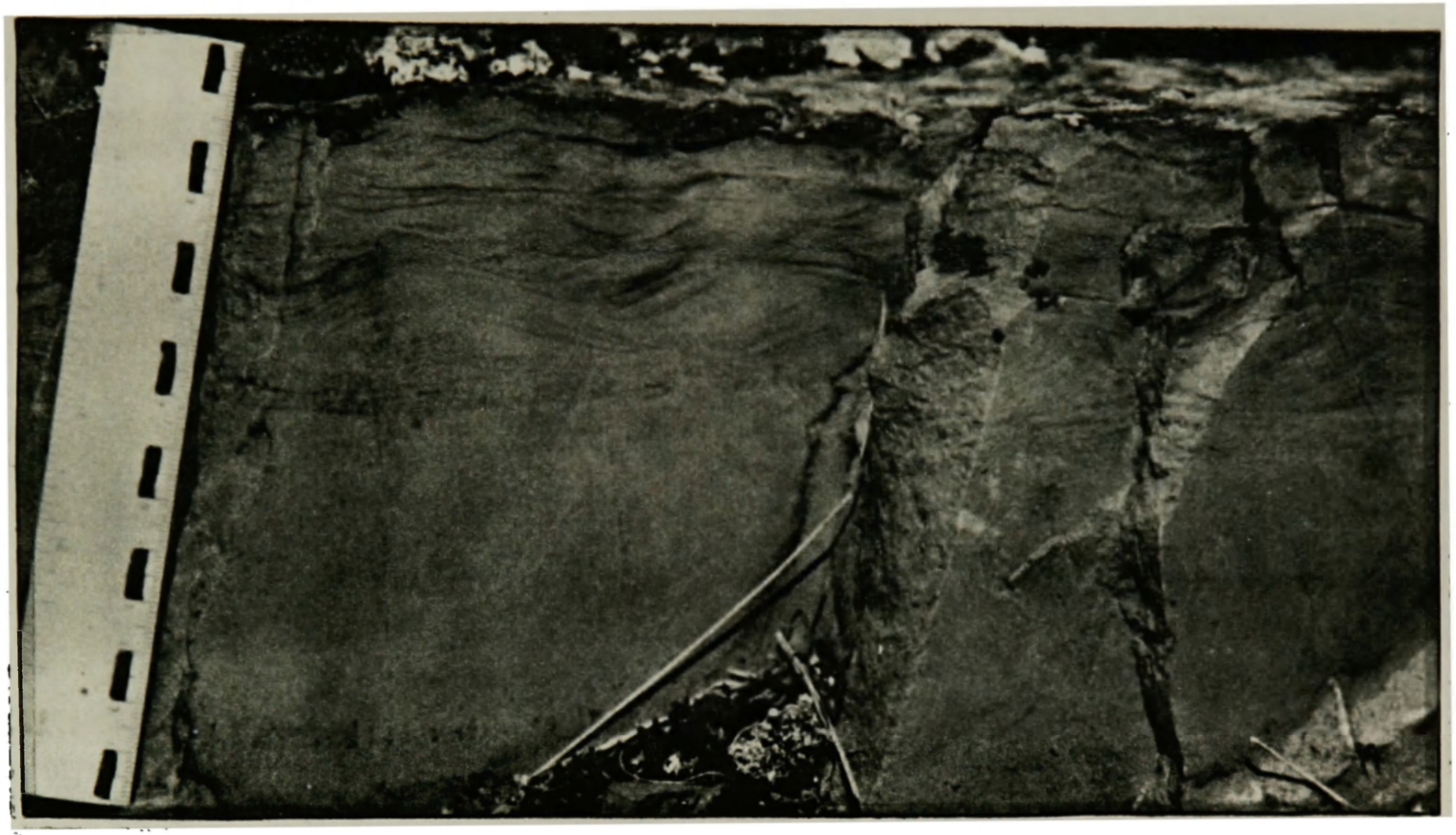


Figure 15. Flat-laminated sand sediment-type. Apparently structureless fine-grained sand overlain by climbing ripples; depositional-stoss ripples pass upward to progressively smaller erosional-stoss ripples.

indicating a depositional contact.

Although flat-laminations characterize this sediment-type, other features include: load structures, mud chips, climbing ripples, shrinkage-cracked muddy tops, and leisigang bands. Load structures at the bases of some flat-laminated sand beds are sunken down as load casts into sand or silty sand below. Flame structures commonly extend into the upper bed bounding each load, and indicate that the redistribution of sediment was upward as well as downward. Some sand pillows thus created are structureless, but others have internal laminations paralleling their external form. Load casts also commonly occur within the lower structureless part of thicker sand beds, or at the interface between their structureless and flat-laminated parts.

Climbing ripple cross-laminations form as much as the upper third of occasional flat-laminated sand beds. In most cases, erosional-stoss climbing ripples (Type A of Jopling and Walker, 1968) pass upward to depositional-stoss climbing ripples (Type B), but in rare instances depositional-stoss climbing ripples pass upward to erosional-stoss climbing ripples, with a concomitant decrease of ripple size (figure 15). Slightly sinuous- to straight-crested asymmetrical ripples capping climbing ripple cosets are occasionally exposed on bedding planes, but straight-crested symmetrical ripples more commonly form the upper bedding surfaces of flat-laminated sand beds. The symmetrical ripples are less than a centimeter high, with wavelengths ranging from one to four centimeters.

Beds of this sediment-type frequently grade up to thin (0.5 - 2 centimeter) mud caps which are cut into well-defined polygons. Some large mud polygons (to 10 centimeters across) are outlined by wide cracks (to 1 centimeter across) and are cut into smaller polygons by one or more orders of progressively narrower cracks. These cracks range from V-shaped to parallel-sided, and they are filled by sediment from the overlying bed. Rounded, matrix-supported mudchips of reworked mud plyygons occur within the lower third of some sand beds.

Leisigang bands, a post-depositional hematite concentration, commonly occur in flat-laminated sand beds. Usually, narrow purple or black leisigang bands accentuate certain flat-laminae. In thinner, silty beds, horizontal leisigang bands often reveal flat-laminations not otherwise visible texturally in the field. Hematite concentrations infrequently create a variety of irregular patches and patterns, including sets of curving lines which resemble cross-bedding, though they cut across flat-laminations. The color of flat-laminated sands ranges from white to grey to purple to red, or uncommonly from green to grey-green.

Most flat-laminated sand beds occur in meter- to tens of meters-scale sets. Three types of vertical sequences occur within bed sets: 1) thinning and fining upward, 2) thickening and coarsening upward, and 3) first thickening, but then thinning upward. Sets of the cross-bedded sand sediment-type occur locally in the thicker, coarser parts of these sequences. Sets grade upward or downward to sets of even couples, or less frequently, pass sharply to sets of even couplets. Uncommonly, isolated flat-laminated sand beds occur within sequences of even couples or even couplets. Based on vertical sequences and stratigraphic correlations, thicker clean sand beds generally pass northeastward to somewhat thinner beds with silty-sand upper portions, which pass in turn to thinner beds wholly composed of silty-sand; these then pass to beds of the even couple sediment-type.

INTERPRETATION

Detailed analysis of its sedimentary features reveals that the flat-laminated sand sediment type was deposited episodically by shallow, terrestrial sheetfloods. This interpretation is based on an analysis of the following considerations: 1) lateral continuity of beds, 2) bed contacts, 3) clean sands, 4) silty-sands, 5) load structures, 6) climbing ripples, 7) symmetrical ripples, 8) muddy caps, 9) shrinkage cracks, 10) color,

11) absence of large-scale cross-beds, and 12) flow regime.

1) Lateral continuity of beds. The great lateral extent and internal continuity of flat-laminated sand beds indicates that they were deposited from unconfined flows across very broad, flat surfaces. This interpretation is confirmed by the absence of scoured surfaces or other discontinuities within beds which characterize channellized flow.

2) Bed contacts. Localization of basal scoured surfaces indicate that in some places the flood moved as local surges which swept and cut the sediment surfaces irregularly. Planar erosional contacts record the passage of unconfined flow capable of uniformly scouring earlier deposits. Non-erosional contacts indicate continual deposition during the flow.

3) Clean sands. Where sands are essentially free of silt and clay, sorting by protracted bedload transport and deposition are confirmed. Flat-laminations can be developed in very fine- to fine-grained clean sands by aggradation of an upper plane bed (Harms and Fahnestock, 1965; Moss et. al, 1980; Allen, 1984b; Cheel and Middleton, 1986) or by the migration of very low relief bedforms under very shallow (< 10 cm) flow at or near upper plane bed conditions (Jopling, 1964; Smith, 1971; McBride et. al, 1975; Moss et. al, 1980; Cheel and

Middleton, 1986). The absence of occasional "micro"-cross-laminations or very small ripples (Lindholm, 1982) and the lack of coarser-grained lenses typical of bedform-deposited laminae (Allen, 1984b) indicate that bedform migration was not a significant depositional mechanism for this sediment type. Instead, upper plane bed conditions are indicated for the flat-laminated sand sediment-type by the lateral extent and evenness of the laminae, and by occasional martrix-supported mud-pebbles (Fahenstock and Haushild, 1962; Moss, 1972).

Most sandstone beds in other rocks which appear massive in the field are found to be flat-laminated or cross-laminated when examined by X-ray or etching techniques (Hamblin 1962a, 1962b). The structureless lower portion of some thicker beds may actually be flat-laminated because, if grains deposited on an upper plane bed are not segregated by size, no horizontal laminations form (Simons et. al, 1965; McBride et. al, 1975). Since the narrow size range from very fine to fine sand is difficult to sort into recognizable laminae, the structureless sand layers probably record higher flow velocities under which laminae aggraded too rapidly to be well sorted. The upward gradation from apparently structureless to flat-laminated clean sands thus records flow deceleration. Some structureless sands may, on the other hand be truly homogenous deposits formed by compaction dewatering during very rapid sedimentation (Reineck and Singh, 1980), still confirming a waining flow interpretation.

4) Silty-sands. Flat-laminated sand with upward-increasing admixtures of silt and small amounts of clay can be formed by two processes: 1) particles settling directly out of suspension in standing water, or 2) suspension sedimentation from flowing water. The deposits of these two processes can be distinguished by differences in the sharpness of the resulting stacked laminae.

Admixtures from direct suspension sedimentation produce sediments in which coarser and finer laminae are juxtaposed (Singh, 1972). Since the flat-laminations of silty sand beds in the flat-laminated sand sediment-type are characteristically subtle and of uniform texture, they are not the product of direct suspension sedimentation.

Instead, the deposition of flat-laminated silty-sand results from suspension sedimentation in current velocities capable of bedload traction, whereby sediments settle into the bedload (Kuenen, 1966b, 1967; Middleton, 1967; Kuenen and Sengupta, 1970; Lowe, 1982; Harms et. al, 1982). The bedload layer is fed by particles settling from a flow whose capacity for retaining suspended load is exceeded by deceleration of the flow (Lowe, 1986). In dropping the excess load, all grain-sizes in suspension are deposited, but the higher settling rate of the coarser grains causes the suspended load to fine progressively as it travels downstream (Kuenen, 1966b, 1967; Kuenen and Sengupta, 1970; Ghosh et. al, 1986). This process produces a distally-fining deposit. Because the flow's leading edge travels across any surface at a higher velocity than does its trailing edge, the initial currents are charged with sediment. But, with deceleration, the flow becomes progressively depleted in coarser grains and thus deposits a fining-upward layer. The flat-lamination of these graded beds is produced by accumulation that is slow enough to sort the grains (Kuenen, 1966a; Banerjee, 1977). Thus, the gradation from clean flat-laminated sand to silty flat-laminated sand records flow deceleration which forced the suspended load to accumulate.

The deposition of bedload sand, silt, and clay together in individual laminae is controlled by hydraulic and packing processes. Hydraulically, pulsations in near-bed turbulence, and the existence of a thin layer of laminar-flow at the bed permit deposition of silt, and even clay, directly onto or into the bed (Simons et al., 1963; Kuenen and sengupta, 1970). However, the packing of grains into a framework from a bedload dominated by grain collisions is equally or more important than their hydraulic selection (Moss, 1972). Therefore, silt is readily accepted into sandy laminae, often to the point that a bimodal sediment is produced (Moss, 1972). Furthermore, clay and fine silt are incorporated into flat-laminae because they are trapped between accumulating grains at a concentration similar to that in the suspended load (Simons et. al, 1963; Kuenen, 1966b; Kunenen and Sengupta, 1970; Moss, 1972).

Some fine silt and clay probably entered flat-laminated sands by the infiltration of slow moving or ponded turbid flood water. Infiltration is characteristic of terrestrial sheetfloods (McGee, 1897; Walker et al., 1978; Turner, 1980; Molenaar, 1986; Flint; 1987). Finally, authigenic clay minerals contribute to the dirty appearance of some flat-laminated sand beds (Herndon, 1983).

5) Load structures. Load structures within the lower structureless parts of thicker sand beds, or at the interface between their structureless and flat-laminated parts result from slowly deposited, densely packed layers deposited from traction, loading down into more rapidly deposited, more loosely packed layers also deposited from traction. Load structures at the bases of flat-laminated sand beds record densely packed flat-laminated sand deposited by traction that loaded down into more loosely

packed lower regime or suspension-with traction deposits (Moss, 1972; Visher and Cunningham, 1981) of the previous depositonal event. In both cases, rapid deposition and loading forced pore fluids to flow up through pillars between the loads, up to the sediment/water interface, liquefying the pillars (Lowe, 1975; Visher and Cunningham, 1981). In some load structures, the loading layer was also liquified by the ascending fluids, and its dewatering and re-packing destroyed primary laminations. In other load structures, confining pressure was sufficient to maintain the flat-laminations of the loading layer, so that the laminations parallel the external form of the load cast (Visher and Cunningham, 1981). The continuity of load deformation up into flat-laminations show that loading occurred during deposition, not as later, large-scale disruption of the sediment package by dewatering and compaction, or seismic shocks (Lowe, 1975; Mills, 1983).

6) Climbing ripples. The common upward transition from erosional-stoss to depositional-stoss climbing ripples records the decreasing bedload traction rate and increasing suspension settling rate of a waining flow. (Jopling and Walking, 1968; Ashely et. al, 1982). The very rare, opposite upward transition from depositional-stoss to erosional-stoss climbing ripples requires that, with time, the suspension settling rate

decreased relative to the bedload transport rate, while the upward decrease in grain- and ripple-size indicates decelerating flow (Jopling and Walker, 1968). This sequence is interpreted to record waining flow which held progressively less sediment in suspension, as discussed in the preceding analysis of silty flat-laminated sands.

7) Symmetrical ripples. The straight-crested symmetrical ripples which cap some flat-laminated sand beds are interpreted to have formed by wave-oscillation in ponded water. The small size and wavelength of the oscillation ripples indicate that they were produced in shallow water (Reineck and Singh, 1980).

8) Muddy cap. The common upward transition from flat-laminated silty-sand to a thin muddy cap records the change from suspension-with-traction to direct suspension sedimentation brought on by flow too slow to transport sediment as bedload, yet still capable of transporting and depositing clay from suspension. Ultimately, the local ponding of the turbid waters allowed the finest suspended matter to settle out in still water.

9) Shrinkage cracks. Although there is some overlap in morphology between subaerially- and subaqueously- or substratally-formed shrinkage cracks (Plummer and Gostin, 1981), those deforming the muddy caps of flat-laminated sand beds consistently possess qualities indicative of desiccation. They are straight

and outline well-defined polygons, and some mud layers are cut by several orders of cracks, recording different generations of cracking, a process unique to desiccation (Picarrd and High, 1973; Reineck and Singh, 1980; Plummer and Gostin, 1981). The upward curvature of many mud-polygons, and the derivation of crack-fill from above also indicate subaerial desiccation. Deep drying of sun-cracked muds produces wide, deep, V-shaped cracks and a considerable shrinkage in sediment volume. In addition, the mud clasts within flat-laminated sand beds are probably derived from dried, durable, ripped-up mud polygons.

Cracks produced by synaeresis, the subaqueous dewatering of clay minerals, are unlike those observed in this study. Synaeresis cracks are reported have a sinuous or spindle plan shape (Donovan and Foster, 1972; Picard and High, 1973; Plummer and Gostin, 1981), and they rarely develop into polygonal patterns (Burst, 1965; Reineck and singh, 1980). Subaqueous and substratal synaeresis cracks do not curl at their edges, and can be filled from below as well as above (Plummer and Gostin, 1981). Synaeresis produces narrow fissures (Burst, 1965; Reineck and Singh, 1980), and thus involves less water-loss and contraction than desiccation does. Compaction of buried synaeresis cracks is likely to widen and crumpel them (Donovan and Foster, 1972), because the moisture content of clays is still relatively high.

10) Coloration. The prevalent grey, purple, and red colors of flat-laminated sand beds (and Ravalli Group rocks in general) are due to hematite disseminated throughout the rocks as pore filling and grain-rims. The hematite, an authigenic mineral and alteration product of detrital goethite, magnetite, or ilmenite, reflects oxidizing diagenetic conditions (Herndon, 1983) characteristic of deposition alternating with subaerial exposure (Turner, 1980; Molenaar, 1986; Flint, 1987). Accordingly, horizontal leisigang bands mostly parallel oxidized heavy mineral laminae, while concentric leisigang patterns probably record diffusion through the loose sediment.

11) Absence of large-scale bedforms. The common vertical transition from upper plane bed lamination to lower regime ripple-bedding without intervening large-scale cross-stratification is contrary to the equilibrium flow regime model (Simons et. al, 1965). Three factors probably contributed to skipping the megaripple stage: 1) the virtual inability of very fine-grained sands to form megaripples (Harms et. al, 1982), 2) disequilibrium between the flow and the bed (Allen, 1973; Allen and Collinson, 1974), and 3) shallowness of the flow (Singh, 1972).

Megaripples would not be a favored bedform simply because very fine-grained sand is the dominant constituent of the flat-laminated sand sediment type. However, since fine-grained sand is locally abundant in these deposits, and the formation of very fine-grained megaripples is possible, although uncommon, this factor does not in itself explain their absence from the vertical transition.

Due to their size, the build-up and maintainence of megaripples require that the proper flow conditions persist for a lengthy time period (Allen, 1973, Allen and Collinson, 1974). Thus, the absence of large-scale cross-stratification may indicate that the unconfined flow decelerated too quicly for megaripples to develop. The high sedimentation rate recorded by load structures and climbing ripples confirm that flow wained rapidly.

A flow depth of several decimeters or more is necessary for megaripple development (Singh, 1972). Since upper plane bed stability requires increasing flow velocity with increasing depth, plane beds are less abundant than lower flow regime bedforms in deeper flows (Harms et. al, 1982). The overwhelming dominance of upper plane laminations in this sediment type therefore indicates that the depositing flow was normally shallow; the abscence of megaripples suggests that the flow was usually less than a few decimeters deep.

Wave-oscillation ripples also indicate shallow water deposition, and mudcracks indicate subsequent exposure of flat-laminated sand beds.

12) Flow regime. The planar bed phase of the flat-laminated sand beds suggests that upper plane bed conditions dominated this system by maintaining a balance between flow depth and velocity through aggradation of the bed. However, the upper portion of many silty sands were deposited under lower regime flow, in which the development of ripples and megaripples was suppresed by the unsteadiness of unconfined flow (Moss, 1972), and the high apparent viscosity of the bedload layer, which dampens turbulence near the bed to the point that bedforms are not stable (Allen and Leeder, 1980). Loading into bed tops suggests that silty-sands frequently possessed the grain-packing of lower flow regime sediments despite their plane bed stratification (Moss, 1972). Occasional climbing ripple sets demonstrate a successful transition from an upper to a lower regime bed phase, and re-affirm the conclusion that silty sands were derived from suspension, but emplaced as bedload.

Conclusion: The preceding considerations indicate that beds of the flat-laminated sand sediment-type were deposited by episodic, shallow, unconfined, dominantly upper regime flows which inundated broad, flat surfaces as continually decelerating sheets of turbid water. The predominance of upper regime flow, the high suspended sediment content, and the great lateral extent of beds indicate that the flows were of a flood magnitude; and conform to the concept of sheetfloods as a "sheet of unconfined flood water moving down a slope" (Hogg, 1982). Intervening mud layers, wave-oscillation ripples, and sun-cracks confirm that the flows were discrete, ephemeral flood events separated by time intervals. Where these features are absent between sand beds, flood waters may have infiltrated quickly without ponding, they may have been scoured by the next sheetflood, or stacked beds were deposited by different pulses of a single flood event (McKee et. al, 1967; Stear, 1985).

The fronts of some sheetfloods were capable of scouring the ground surface, but as flow inundated a progressively larger area, it thinned and decelerated, forcing sediment to deposit continually. With each flood, a water mass surged out onto a sandy plain, becoming the highest topopraphic feature in the area. The resulting hydraulic head caused the fluid to spread out evenly in many directions, waining continually as it flowed, and depositing the vertically and laterally fining beds of the flat-laminated sand sediment type. The term sandflat (Hardie et al., 1978) describes the gently sloping sandy apron deposited by sheetfloods distal from braided channels.

Sheetfloods have been observed in recent times (McGee, 1897; Rahn, 1967), but their ephemeral and catastrophic nature leaves the study of recent (McKee et. al, 1967; Williams, 1971; Tunbridge, 1983; Blair, 1987)) and ancient sheetflood deposits (Hardie et al, 1978; Tunbridge 1981, 1984; Hubert and Hyde, 1982; Sneh, 1983; Smoot, 1983; Cheadle, 1986; Demicco and Kordesch, 1986; Flint, 1987) as the only practical means to discover their depositional processes. Modern and ancient sheetflood deposits are characteristically similar to those of the flat-laminated sand sediment type. Miall (1985) recognizes sheetfloods to be part of the spectrum of fluvial systems.

EVEN COUPLE SEDIMENT-TYPE

DESCRIPTION

The even couple sediment-type is characterized by 3 to 10 cm thick, laterally extensive, even, graded beds which contain sandy or silty lower layers, and muddy upper layers (figures 16 and 17). Even couples can often be traced across 30 meters of outcrop without thinning, and appear tabular in smaller outcrops. The lower layers fine-upward, consist of very fine-grained silty-sand, sandy-silt, or silt, and commonly include matrix-supported mudchips. Rarely, mudchips are concentrated at the bases of even couples, forming layers

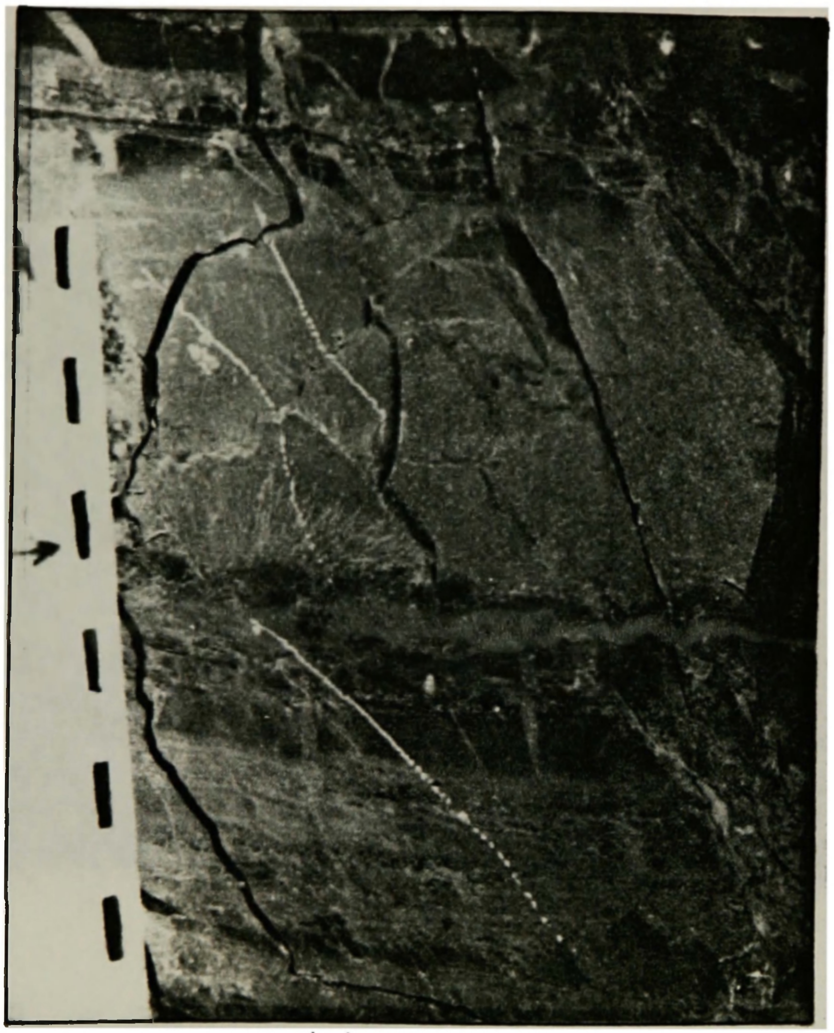


Figure 16. Couple sediment-type. Couples with cracked clay layers. Mudchip conglomerate layer (center) forms the base of a thick couple with a graded silt layer and a thin, sun-cracked clay cap.

Figure 17. Couple sediment-type. Couples with flat-laminated sandy-silt layers and thin, sun-cracked clay layers. Bands of diagenetic hematite concentrations accentuate flat-laminations.

of intraformational conglomerate (see figure 16). Flat-laminations characterize the silty layers (see figure 17), but some are simply graded, while others, less common ones, are ripple cross-laminated (see figure 16). The upper muddy layers are continuations of the fining-upward silt below, but they stand out as distinct layers because the fine silt and clay content of the sediment increases dramatically. Symmetrical ripples occasionally cap the silty layer and have a sharp boundary with the overlying muddy layer. The muddy layers are generally less than a third of an even couple's thickness, and are commonly mudcracked (figure 16 and 17). The cracks are similar to those described with the flat-laminated sand sediment-type and are interpreted to be desiccation cracks for the reasons cited for the flat-laminated sand mudcracks.

Bedding contacts bounding even couples are sharply marked. Most contacts are planar, and determining whether one is a depositional surface or a plane of minor scour is often difficult. Where mud polygons of a cracked layer are dislocated, a scoured contact is confirmed.

The distinction between the coarser and finer layers of even couples is accentuated by the differing color or shade of their silty and muddy layers. In terms of the coarser-to-finer coupling, colors range from grey-to-purple to lighter and darker shades of purple or red, and uncommonly, green. Generally, the silty layer of even couples are lighter and paler, grading upward to more brillant hematitic (purple or red) or chloritic (green) muddy layers.

Even couples usually form meters-scale sets which pass vertically to sets of couplets or flat-laminated sand beds, but occasional, isolated even couples do occur within sets of these two sediment-types. Based on vertical sequences and stratigraphic correlations (see chapter on stratigraphy), even couples thicken and coarsen to the south and west, and thin and fine to the north and east. Similarly, even couples pass southward and westward to flat-laminated sand beds, and to couplets northward and eastward.

INTERPRETATION

Both even couples and thin, silty flat-laminated sand beds contain a fining-upward layers composed of flat-laminae, but even couples are thinner, finer-grained, and have a noticeably thicker muddy caps. These similaritities indicate that their depositional processes were similar, while these differences show that even couples are the product of more distal sedimentation. Accordingly, the silty layer of a even couple is interpreted as an upper flow regime plane bed aggraded by suspension-with-traction sedimentation within a shallow, waining sheetflood overloaded with suspended sediment (see discussion of the flat-laminated sand sediment-type). The ripped-up and transported mudchips attest to the vigor of the sheetfloods' early stages. As in flat-laminated sand beds, ripple cross-laminations record a shift to lower flow regime conditions.

The abrupt increase in clay deposition and the suspension sedimentation of the mud layer indicate a sharp drop in current strength followed by ponding, as water rose and progressively inundated the alluvial sandflats. Mudcracked even couples indicate that the shallow body of water resulting from the sheetfloods eventually receded and exposed the sandflats to desiccation. Subsequent sheetfloods often swept these broken surfaces, ripping-up mud polygons, and transporting them a short distance. The mudchips were deposited early on, towards the base of beds, because the flood's rapid expansion across the very flat surface caused the flow to thin and decelerate continually. Flow capable of scouring and transporting mudchips proximally was thus forced to deposit them distally.

In even couples with symmetrical ripples at the top of the silty layer, waves in the standing water left after flooding impinged on the sediment surface, winnowing and reworking sediment into oscillation ripples. Eventually, calmer conditions allowed the mud to settle out of the shallow, turbid water, depositing the sharply bounded, overlying mud layer. Through infiltration and evaporation, the water disappeared, leaving a dry surface which was commonly sun-cracked.

COUPLET SEDIMENT-TYPE

DESCRIPTION

The couplet sediment-type in the Ravalli Group is characterized by 0.3 to 3 cm thick, laterally extensive, sharply bounded, fining-upward beds composed of a lower flat-laminated, ripple cross-laminated, or simply graded silty layer gradationally or sharply overlain by a clayey upper layer (see figures 18, 19, 20). The lower layers range from very fine-grained silty sand, to silt, to clayey silt, and commonly contain rounded, matrix-supported mudchips. Cracks commonly cut the upper clayey layers into mud polygons. Where mud polygons are dislocated, or the lower silty layers contain ripped-up mudchips, local basal scour is recorded, but most couplet bases show no evidence of being scoured. Couplets are tabular and continuous in small exposures, and individual couplets can often be traced across outcrops 10 meters wide.

Both the lenticular and even couplet sediment-types defined by Winston (1986c) are included in the couplet sediment-type as recognized here because, although differential weathering and cleavage refraction make

Figure 18. Couplet sediment-type. Even and lenticular chloritic couplets. Note oscillation ripples capping the thick silt layers.

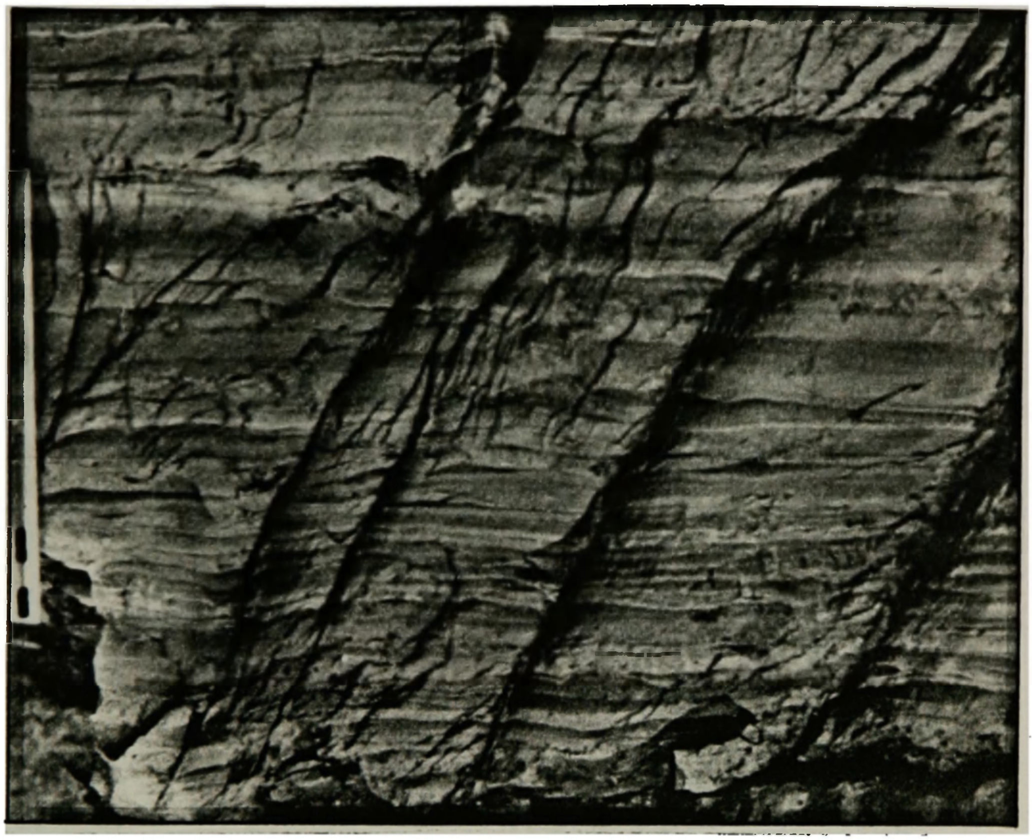


Figure 19. Couplet sediment-type. Even and lenticular chloritic couplets. Note darker symmetrical ripples overlain by lighter mud.

Figure 20. Couplet sediment type and mudchip conglomerate sediment-type, sub-type 1 (top of photo to the right). The base of this boulder consists of hematitic, mudcracked even couplets overlain by a bed of sub-type 1 mudchip conglomerate. Spindle-shaped, sediment-filled cracks of varying veritical extents are widest in the stratum composed of stacks of curled even couplets, and die-out in the overlying set of flat-lying, mudcracked even couplets.

couplets evident in outcrop, cleavage and lichens often obscure the details of internal layering, making it difficult to consistently determine whether the lower silty layers are of constant thickness, as in even couplets (figure 18, 20), or comprise the straight-crested symmetrical ripples of lenticular couplets (figure 18, 19). Thus, the couplet sediment-type as used here is similar to the flat-laminated sand and couple sediment-types in that some beds include symmetrical ripples, and others do not.

A distinctive sub-variety of even couplets recognized in this study is termed the alpha couplet variety. Alpha couplets are continuously graded and rich in clay from base to top, so their upper and lower layers are less distinct than those of regular even couplets. The lower silty layers of alpha couplets do not contain flat-laminations, cross-laminations, or mudchips, and their upper layers are not mudcracked. Occasional strata up to 4 centimeters thick are composed of broadly U-shaped pillows of concentrically deformed alpha couplets separated by structureless mud pillars.

Lenticular and even couplets range from grey to purple to red, and less commonly they are green. Very rarely, green silty layers are overlain by purple clayey layers. All alpha couplets are green. Generally, the lower silty layers of couplets are lighter in color, and grade up to the darker colored clayey layers (see figures 18, 20), but darker-to-lighter coupleting occurs locally (figure 19).

Red, purple, and grey couplets are hematitic, and green couplets are chloritic. The hematite and chlorite are diagenetic minerals (Herndon, 1983) formed under oxidizing and reducing conditions, respectively, so the color of couplets is a product of diagenesis. However, lenticular couplets are most commonly chloritic, while even couplets are most commonly hematitic, and hematitic couplets are more commonly mudcracked than chloritic couplets.

Lenticular and even couplets usually form meter- to tens of meters-scale sets which pass vertically to sets of couples or, uncommonly, to sets of flat-laminated sand beds. Occasional isolated couplets occur within sets of couples or flat-laminated sand beds. Thickly coupleted intervals in the Grinnell Formation of the eastern side of the study area pass westward to intervals of even couples (see figure 28). Lenticular and even couplets are generally thicker and siltier toward the west, and thinner and muddier toward the east.

Alpha couplets form distinctive, sharply-bounded sets up to a meter thick. The sets of alpha couplets and the distinctive underlying and overlying sandy sediments are easily recognized in outcrop, and are used as keybeds in the stratigraphic correlations (figure 28, Appendix F). One keybed occurs in the upper Revett Formation of North Crow Creek; another correlates between the St. Regis Formation of North Crow Creek and the correlative part of the Grinnell Formation at Lion Creek (see figure 28, Appendix F). The only easily accessible outcrops of a stratigraphic keybed are near the base of Lion Creek sub-section 4 (Appendix C).

In outcrop, a well-developed, closely spaced cleavage penetrates the entire set of alpha couplets at a high angle, and is quite distinct from the much less continuous cleavage in even and lenticular couplets. The penetrative cleavage reflects alpha couplets' high clay-content and uniform stratification.

The sandy sediments which bound sets of alpha couplets above and below are characteristically green. In the upper keybed at North Crow Creek, a few decimeters of structureless, green, very fine-grained silty sand above a set of alpha couplets is sharply overlain by a bed of structureless, grey, very fine-grained silty sand; deep, broad loads of the grey sand are seperated by narrow flame structures of the green sand. In the lower keybed at North Crow Creek and the one at Lion Creek, the green color of the mudcracked even couples and flat-laminated sand beds bounding sets of alpha couplets fades away from the sets and passes to mottled green and red, and then to just red within two meters. In the mottled areas, the segments of cracked clay layers are often red with green rims. In addition, in the lower North Crow Creek keybed, a decimeter-thick bed of green coarse sand directly overlying the set of alpha couplets includes angular, white, pebble-sized mudchips, and the mudcracked layers of the first few overlying green even couples are also white.

INTERPRETATION

Overview Couplets are interpreted to be the most distal products of the sheetfloods which deposited flat-laminated sand beds and couples on the alluvial aprons and sandflats (see figure 32). As the floods crossed onto the dried mudflats beyond the sandflats, they thinned and decelerated, depositing the lower silty layers of even couplets. Clay settling out of ponded floodwaters formed the upper clayey layers of even couplets. Where the shallow standing waters persisted for some time, wave-currents reworked the silt of even couplets into the symmetrical ripples which characterize lenticular couplets, and their upper clayey layers settled out as the turbulence subsided. Where the sheetfloods flowed into perennial lakes, their sediment load continued traveling as suspension clouds and the silt and clay that settled from the muddy standing water formed alpha couplets. Support for these interpretations comes from even couplets, lenticular couplets, and alpha couplets.

Even couplets In circular flume studies of suspensions of silty sediment similar to that in the couplet sediment type, Banerjee (1977) found that: 1) very rapid deceleration of a silty suspension produces a non-laminated graded silt layer overlain by a clay suspension cap; 2) less rapid deceleration deposits a flat-laminated graded silt layer with a suspension sediment cap; and 3) moderate to slow deceleration yields a graded layer containing ripple cross-laminated silt capped by a clay suspension blanket. Both the non-laminated and flat-laminated graded silt layers are aggraded upper plane beds, while the ripples record aggradation in lower regime flow (Banerjee, 1977).

Although Banerjee's (1977) study was undertaken to model turbidity flows, the processes accompanying deceleration in a circular flume (Kuenen, 1966; Banerjee, 1977) are also analagous to those operating in sheetfloods (see the flat-laminated sand sediment-type). Cummins (1956) first recognized that the similar depositional processes of turbidity flows and sheetfloods produce similar deposits. Although both turbidity currents and sheetfloods produce even couplets, those discussed here are interpreted as distal sheetflood deposits because sun-cracks show that the sediment surface was often subaerially exposed, and because of their distal position relative to flat-laminated sands and even couples, which are interpreted as more proximal sheetflood deposits.

The abundance of sun-cracks and mudchip inclusions in even couplets indicates that the thinned, distal parts of dominantly upper regime sheetfloods scoured dry mudflats, ripping-up mud polygons, and depositing them in fining-upward silty layers as flow wained. The abrupt upward increase in clay content records the transition from flowing floodwater capable of traction transport, to standing water that deposited suspended sediment directly onto the bed. Stated another way, the transition records the expansion of a shallow water body across the mudflats (Smoot, 1983). As the ponded waters infiltrated and evaporated between floods, the mudflats were exposed to desiccation.

Smoot (1985) studied mudcracking processes on modern playa mudflats in Nevada and California, and found that the 0.5-4.0 centimeter-thick graded silty-mud or laminated muddy-silt beds deposited at mudflat margins exhibit angular, open cracks, which are completely filled by sediment in the next flood event. The mudcracks only penetrate a single clay layer because the underlying sediments are not saturated by flood waters (Smoot, 1985). The vast majority of desiccated even couplets (and couples) are cracked in this manner (see the mudcracked even couplets at the bottom of figure 20), suggesting that they formed through similar processes.

Distally, and between sources of sediment influx, open mudcracks trap waters flooding modern playas (Smoot, 1985). Since the flood waters carry only a small amount of sediment, the cracks are usually only partially filled with sediment, while the rest of the surface receives a thin layer of sediment. Occasional floods erode and broaden mudcracks, but most are modified after flooding as water saturates the polygon edges and causes them to slump or flow. Drainage rills develop across polygon edges as water drains from the polygon surface by seeping through the cracks into the underlying sediment. Because the partially-filled crack depressions host the only well-saturated sediments, desiccation cracks repeatedly develop in the same polygonal pattern. Where underlying beds become saturated, this pattern penetrates earlier deposits (Smoot, 1985). These processes form rounded stacks of mud polygons separated by U-shaped troughs which characterize parts of slowly aggrading playa surfaces (Smoot, 1985). In cross-section, the mudcracks are often spindle-shaped and extend to different depths and heights. Smoot (1985) believes that many of the desiccation fabrics he observed are commonly interpreted as dewatering structures.

The stratum composed of stacks of curled, even couplets in figure 20 is interpreted to represent stacks of mud polygons separated by sediment-filled sun-cracks, and to have formed by processes similar to those described above. The unusually thin scale of coupleting in the polygon stratum confirms that this part of the mudflat received only minor influxes of sediment for many floods, while the abundance of sun-cracked even couplets throughout figure 20 indicates that the mudflat was frequently desiccated. The fact that the cracks are widest in the mud polygon stratum records that the sides of the polygons stacks flowed and slumped into open cracks. Their spindle shape and differing vertical dimensions indicate that the mudcracks penetrated underlying saturated sediments and influenced cracking in overlying couplets to varying extents, and prove that they are not dikes formed by explosive dewatering of the sub-type 1 mudchip conglomerate bed. Instead, the conglomerate bed may have been produced in-situ by the repeated development of different polygonal patterns on a slowly or non-aggrading mudflat, or it may be a mudflow deposit (see sub-type 1 of the mudchip conglomerate sediment-type). Nevertheless, the possibility that water moved upward through the cracks at times is not discounted, since playa mudcracks often act as conduits for occasionally or seasonally rising groundwaters (Eugster and Hardie, 1978; Muir et al., 1978).

<u>Lenticular couplets</u> The symmetrical profile and straight crests of the ripples which characterize lenticular couplets indicates that they were developed by oscillatory (wave) currents (Harms et al., 1982). Given their interbedding with even couplets, lenticular couplets are interpreted to have been deposited when the sediment transported onto the mudflats by a sheetflood remained submerged long enough for wind-generated waves to rework the sediment deposited as even couplets into oscillation ripples. Where silty sediment was abundant, the symmetrical ripple aggraded (figure 18); where silt was scarce, unconnected, "starved" ripple lenses formed (figure 19). Periodically, the water quieted to the point that suspended clay settled-out, forming the capping clay layer. Occasional sun-cracked lenticular couplets evidence times when the mudflat was exposed long enough to be thoroughly dried and desiccated.

<u>Alpha couplets</u> Their high clay content and simply graded internal structure indicate that alpha couplets were deposited as sediment settled from turbid standing water directly onto the bed. The indistinct coupleting thus records the gradual exhaustion of the suspension's fine silt supply, such that only slower-settling clay particles remained to form the capping layer. The

absence of flat-laminations, ripple cross-laminations, and transported mudchips confirms that alpha couplets were deposited in still water. The lack of oscillation ripples suggests they were deposited below fair weather wave base, possibly as much as a few meters deep. The reduced state of alpha couplets, evidenced by their green color, is consistent with deposition in a stagnant water body with an anoxic bottom layer and/or substrate.

Deformed sets of alpha couplets indicate that the sediment surface sank to form pillow structures and displaced saturated mud below which rose upward as pillars. They show that the sediment remained "soupy" for extended periods of time, probably reflecting standing water, also corroborated by the absence of mudcracked alpha couplets.

Because the sets of alpha couplets are intercalated with even couples and flat-laminated sand beds deposited on desiccated sandflats or mudflats, the water body is interpreted to have been a perennial lake. Accordingly, where the sheetfloods which deposited beds of the flat-laminated sand, even couple, and even couplet sediment-types flowed into a perennial lake, their sediment load continued traveling as suspension clouds which muddied the standing water. Settling fine silt and clay formed the lower layers of alpha couplets, while the slowest settling clay particles became the upper layers.

That the perennial lakes were surrounded by desiccated sandflats and mudflats that were washed by ephemeral sheetfloods suggests that the lakes were sustained by springs. That alpha couplet sets are so thin (< 1 meter) indicates that the perennial lakes were relatively short-lived. The scarcity of alpha couplet sets indicates that perennial lakes were uncommon on the Ravalli Group depositional landscape, but the fact that one stratigraphic keybed extends from North Crow Creek to Lion Creek, presently a distance of 15 miles, indicates that when perennial lakes did exist, they could be quite large.

The large loads of grey silty sand in the structureless green silty sand above the set of alpha couplets in the upper North Crow Creek keybed indicate that rapid deposition of the grey sand bed induced abrupt compaction-dewatering of the lower sand bed, since rising rising pore fluids liquefied sediment that rose to form flame structures, while the overlying bed sank as load structures (see the discussion of load structures in the flat-laminated sand sediment-type).

Fading of the green coloration away from sets of alpha couplets, and the passage to mottled green and red, and them red sediments suggests that oxygen-poor fluids from the alpha couplets diffused into the surrounding oxidized sediments and reduced them. The red mud polygons with green rims in the mottled areas support this interpretation. Accordingly, the white mudchips in the coarse sand bed and the white, mudcracked clay layers setting above the alpha couplet set in the lower keybed at North Crow Creek record bleaching near the set of alpha couplets, where the reducing action was most intense.

MUDCHIP CONGLOMERATE SEDIMENT-TYPE

DESCRIPTION

The mudchip conglomerate sediment-type occurs as 2 to 10 centimeter thick, hematitic, tabular beds characterized by abundant sub-rounded to rounded mudchips. Three sub-types of mudchip conglomerate are recognized.

Sub-type 1 conglomerate beds (figure 20) consist of poorly sorted, randomly oriented mudchips supported in a mud matrix. Scattered grains of the Ravalli Group's medium- to coarse-grained sand population occur in some beds. The conglomerate beds range from being dominated by mudchips, to being mostly mud with only scattered mudchips.

Sub-type 2 conglomerate beds contain moderately well-sorted, flat-lying and imbricate mudchips supported by a silty-mud matrix (figures 21 and 22). Uncommonly, thin (1 - 2 centimeter), cracked mud layers sharply

Figure 21. Mudchip conglomerate sediment-type, sub-type 2. Note that the upper mud layer is brecciated in-situ and the lower cracked mud layer is scoured away to either side of the photograph.

Figure 22. Mudchip conglomerate sediment-type, sub-type 2 and coarse sand sediment-type, bedded sub-type. Note the remnant of a sun-cracked mud layer (lower right), and the coarse sand bed's irregularly scoured base.

overlie sub-type 2 conglomerate beds (figure 21).

In sub-type 3 conglomerates, well-sorted mudchips are clast-supported, with little or no mud matrix. Most of the mudchips are equi-dimensional, so no sedimentary structures are apparent within beds, but some beds include occasional flat-lying or imbricate, platy mudchips.

The mudchip conglomerate sediment-type is distinct from the mudchip layers which occur at the base of some couples and couplets in that mudchip conglomerates are not part of a graded stratum, and are not themselves graded.

In small outcrops, beds of the mudchip conglomerate sediment-type appear tabular, with planar basal contacts, and in some places extend across outcrops a few meters wide. Most mudchip beds are interbedded with thin (0.5 to 2 centimeter), muddy, hematitic even couplets, which are commonly sun-cracked. Occasionally, beds of the mudchip conglomerate sediment-type overlie one another directly, forming sets (figure 22).

Mudchip conglomerates are volumetrically a minor constituent of the measured sections and each subtype is restricted to particular stratigraphic units. Most sub-type 1 conglomerates occur in the part of the Grinnell Formation that correlates with the St. Regis Formation of the Lion Creek, Blaine Mountain, and Deep

Creek sections, but they also occur at Blaine Mountain in the unit that correlates with the Upper Revett Formation. Sub-type 2 conglomerate beds occur only in the redbed interval low in the Empire Formation at Blaine Mountain, and sub-type 3 conglomerate beds occur only in the lowest redbed intervals in the Empire Formation at Lion Creek (Appendix E).

Mudchip beds are usually poorly exposed because they are closely cleaved. Their crumbly outcrop and the fact that mudchip conglomerates occur in the highest, most inaccessible parts of the measured sections makes their detection and careful study difficult. As a result, mudchip conglomerate beds may be more abundant in the units in which they occur than is depicted in the stratigraphic sections (Appendix D).

INTERPRETATION

Overview. Some sub-type 1 mudchip conglomerate beds probably record cohesive debris flows, while others are probably desiccation conglomerates produced in-situ; sub-type 2 conglomerates are sheetflood deposits; and sub-type 3 conglomerates are interpreted as lacustrine beaches or shoals washed by waves.

Sub-type 1. Some sub-type 1 mudchip conglomerates probably record sediment-charged, distal sheetfloods that scoured mud polygons from the desiccated mudflats, attaining the consistency of Lowe's (1982) type A cohesive debris flows. These are a form of sediment gravity flow in which clasts are supported by the buoyancy and cohesiveness of their clay-water matrix. As little as five percent clay in the flow volume provides buoyant uplift, reducing the effective weight of the clasts, and lubricates grains, preventing frictional locking (Lowe, 1982). Type A cohesive debris flows congeal as flow decelerates, forming structureless, matrix-supported, pebbly mudstone (Lowe, 1982) similar to sub-type 1 mudchip conglomerate.

Other sub-type 1 mudchip conglomerates were probably produced in place by repeated wetting and drying on slowly aggrading mudflat. In studies of modern playa mudflats, Smoot and Katz (1985) found that "where flood waters pond and the sediment are uniformly saturated, new polygonal crack patterns randomly cross-cut the old." "The mudcracks have interpenetrating fillings indicative of recracking of mudcrack fills, and complex, branching and sinuous cross-sections produced by superimposed crack patterns and soft sediment flowage." The repetition of mudcracking processes "destroys any pre-existing layering, leaving only isolated fragments surrounded by crack fillings. These deposits are soil-like in character and reflect the conditions of low sediment accumulation rates." (Smoot and Katz, 1982). Similar desiccation conglomerates are reported from playas in the Coorong Region of Australia (Muir et al., 1978).

Interestingly, some parts of the mud polygon stratum in figure 20 interpreted to be polygon stacks separated by desiccation crack-fillings (see the couplet sediment-type) are similar to the sub-type 1 conglomerate bed, suggesting that parts of the cracked stratum were also brecciated by repeated cracking and filling.

Some sub-type 1 mudchip conglomerate beds contain floating coarse sand grains transported by sheetfloods.

Sub-type 2. Sub-type 2 conglomerates are interpreted to record unusually turbulent distal sheetfloods that flowed across desiccated mudflats. Accordingly, the muddy flows scoured the cracked surface, entraining mudchips into bedload and taking additional silt and clay into suspension. As the flows spread out across the mudflats, turbulence diminished and the overloaded suspended sediment settled out rapidly, incorporating the bedload and bring the flow to a hault. The matrix-supported, flat-lying and imbricate clasts of sub-type 2 conglomerates indicate flow was in the uppeer regime when it rapidly ceased (Moss, 1972; Fahenstock and Hasushild, 1962).

Where turbid water expelled from the flow ponded, a thin mud layer settled atop the sheetflood deposits. After the water dried-up, the mud layer was sun-cracked. Frequently, the mud layer was scoured-up by the next sheetflood (figures 21, 22), contributing clasts to the next conglomerate bed.

Sneh (1983) reports that mudchip debris beds are commonly formed on the terminal floodplain of the ephmeral sheetflood system draining one-third of the Sinai Peninsula.

Sub-type 3. Sub-type 3 conglomerates are interpreted to have deposited on wave-washed lacustrine shoals or beaches, where small breaking waves eroded mud-polygons from the inundated, desiccated mudflats. Oscillation of breaking waves transported the clasts beachward, winnowing away free silt and mud, and breaking and rounding most clasts to equi-dimensional shapes. Occasional, larger, flat-lying and imbricate clasts lie flat on aggraded shoals or are gently imbricate on foreshore slopes of prograding small beaches.

Beds of well-sorted, clast-supported beach conglomerate occur in Triassic lacustrine deposits of South Wales (Tucker, 1978). Eugster and Hardie (1975) interpret 5 to 20 centimeter thick mudchip conglomerate beds as transgressive lag deposits formed when a very shallow lake expanded over an exposed mudflat, reworking mud-polygons by wave erosion.

COARSE SAND SEDIMENT-TYPE

DESCRIPTION

The coarse sand sediment-type comprises laterally continuous, 2 to 10 centimeter thick beds of well-sorted, rounded to well-rounded, slightly feldspathic (5 to 10% feldspar), medium- to coarse-grained sand (figure 9). Matrix-supported mudchips are a common, minor constituent. The beds have no internal reactivation surfaces, can often be traced across outcrops 30 meters thick, and rarely pinch-out on an outcrop scale. Most coarse sand beds have planar, depositional or slightly scoured lower contacts (figure 23), but rare basal surfaces are cut and scoured (figures 22, 24).

The coarse sand sediment-type is sub-divided into a bedded sub-type and a cross-laminated sub-type. Most coarse sand beds of the bedded sub-type appear structureless, due to their uniform grain-size and the absence of heavy mineral laminae, but in some beds of the bedded sub-type, flat-laminations are revealed by subtle grain-size variations. Most coarse sand beds are of the bedded sub-type. Coarse sand beds in which cross-laminae can be discerned are placed in the cross-laminated sub-type. Many cross-laminae are poorly defined, but are visible because of subtle grain-size variations, or by inclined mudchips resting on foresets (figure 24).

Coarse sand beds of the bedded sub-type range from

Figure 23. Coarse sand sediment-type, bedded sub-type. Note the accreted mudballs in, and the irregularly scoured contact between, the lowest two beds; also, sun-cracked mud layers (towards top of photo).

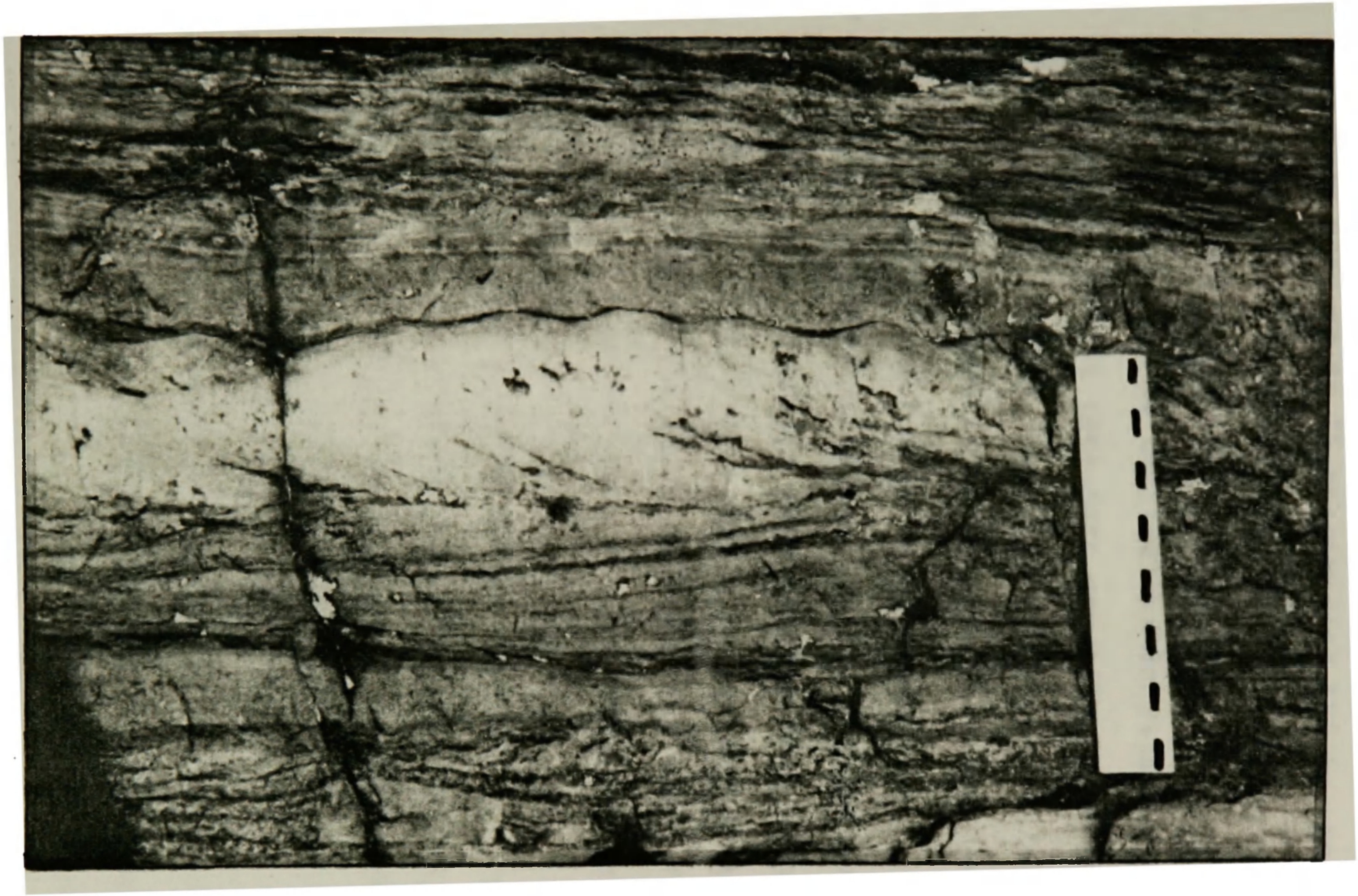


Figure 24. Coarse sand sediment-type, cross-bedded sub-type. Note mudchips resting on foresets (right center) and topset current ripples.

less than one, up to 15 centimeters thick. Most include flat-lying or imbricate, angular to rounded, matrix-supported mudchips (figures 22, 23). Accreted mudballs from 3 to 10 centimeters in diameter are an interesting, albeit very rare constituent of bedded coarse sands (figure 23). The balls consist of a mudchip nuclei with concentrically accreted mudchips and scattered medium to coarse sand grains.

The cross-laminated coarse sand sub-type occurs as solitary, planar-tabular sets ranging from 3 to 15 centimeters thick (figure 10). Cross-laminae are generally high angle (> 20 degrees) and range from straight with angular toes, to straight with tangential toes, to concave.

Coarse sand beds of either sub-type are very rarely topped by sinuous-crested, asymmetrical ripples. Uncommonly, one or two centimeter-thick mudcracked mud drapes sharply overlie coarse sand beds (figure 23).

Most beds of the coarse sand sediment-type are tightly quartz cemented and white, although minute amounts of hematite or chlorite impart a purple, red or green tint to some beds. Rare, medium- to coarse-grained sand beds are carbonate cemented and weather tan or light brown.

Coarse sand beds are usually isolated within sets of couplets; less frequently within sets of couples; and rarely within sets of the flat-laminated sand sediment-type. Occasionally, stacked beds of the bedded coarse sand sub-type form decimeters- to meters-thick sets (figure 23), but at Deep Creek in the St. Regis correlative interval of the Grinnell Formation they form tens-of-meters thick sequences (figure 28, Appendix E or F).

The numbers of coarse sand beds decrease from east to west in the the Burke correlative Grinnell interval, the Revett Formation and its correlative Grinnell interval, the St. Regis Formation and its correlative Grinnell interval (figure 28, Appendix E or F). Beds of the coarse sand sediment-type occur in Unit A, and in Unit B only at Deep Creek (Appendix E or F).

INTERPRETATION

Overview. Coarse sand beds of the bedded and cross-laminated sub-types are interpreted to be deposits of sheetfloods from a different source than those which deposited beds of the flat-laminated sand, even couple, and couplet sediment-types. Beds of the bedded coarse sand sub-type are aggraded upper plane beds, while beds of the cross-laminated coarse sand sub-type record the migration of a slipface at the downstream edge of depositional sheets, or isolated bars within the sheetfloods.

Bedded sub-type Coarse-grained flat-laminated sand beds with flat-lying and imbricate, matrix supported mudchips are interpreted as upper regime plane bed deposits. The apparently structureless coarse sand beds with matrix-supported mudchips are also interpreted as aggraded upper plane beds which were deposited so rapidly that time did not permit sorting of the uniform bedload involved into well-formed laminae (Simons et al., 1965; McBride et al., 1975). The absence of heavy mineral laminae, which so often accentuate laminations in the even couple and flat-laminated sand sediment-types, makes structures within coarse sand beds especially difficult to see in the field.

The great lateral extent and internal continuity of medium- to coarse-grained sand beds indicates that they are sheet sands deposited over broad, flat surfaces during discrete depositional events. Because coarse sand beds of both sub-types intercalate with sun-cracked sediments and are occasionally directly overlain by desiccated mud drapes, the unconfined flows are interpreted as terrestrial sheetfloods.

The mutally exclusive stratigraphic relationship between the Ravalli Group's medium- to coarse-grained and very fine- to fine-grained sand populations (see the "Overview" section of Chapter 2, Sediment-Types). indicates that they are not simply sorting products of a single sand population. Instead, the westward decrease in the numbers of coarse sand beds, and their restriction to the eastern margin of the basin in Unit A and Unit B indicates that the coarse sand population was derived from the east, while eastward-thinning and -fining stratigraphic intervals and beds of the fine-grained sand population (represented in the cross-bedded sand, flat-laminated sand, couple, and couplet sediment-types) indicate that the fine sand population was derived from the west or southwest. Paleocurrent studies of the cross-laminated coarse sand sub-type (below) and the cross-bedded sand sediment-type at North Crow Creek confirm these interpretations, since the direction of net sediment transport for the coarse sands was westerly and for the fine sands, northeasterly. Accordingly, unconfined floods charged with a bedload of medium- to coarse-grained sand from an eastern source surged onto, and spread out across vast flats composed predominantly of fine-grained sediment (couplets, couples, and thinner beds of the flat-laminated sand sediment-type). Most of these sheetfloods flooded only the eastern margin of the mudflats before they had spread out to their full extent, so coarse sand beds dominantly occur interbedded with the most distal products of the western sheetfloods, couplets (figure 32). However, occasional large eastern sheetfloods spread clear across the vast Ravalli Group plain, inundating the mudflats and then flowing even farther west to cover the sandflats. The coarse sand beds thus deposited on the sandflats became interbedded with couples or thin flat-laminated sand beds.

The sheetfloods scoured mud polygons and incorporated them into the bedload. Rounded, pasty mudclasts rolling along the bed gathered mudchips and sand grains, forming accreted mudballs. Each flood thinned and decelerated as it spread out across the flats, and as a result of the widespread reduction in flow competence, the sediment was rapidly deposited over a broad area. Because of the simultaneous thinning, slowing, and bed aggradation, upper flow regime conditions were maintained even as the flow wained. The scarcity of ripples atop coarse sand beds suggests that lower flow regime conditions seldom existed, but some coarse sand was probably too coarse to form ripples (Harms et al., 1982).

The existence of upper plane bed conditions throughout sheetflood events suggests that their waters were generally rather shallow, since deep waining flows normally form megaripples, or at least ripples atop their earlier upper plane bed deposits (Williams, 1971). While it is true that, due to their large size, megaripples require the proper flow conditions to persist for some time in order to develop (Allen, 1973; Allen and Collinson, 1974), it is unlikely that the sheetfloods always wained too rapidly for these bedforms to grow. Thus, the absence of trough cross-bedding from the upper part of coarse sand beds suggests that the falling stages were often less than a meter deep. Similarly, although ripple generation may have sometimes been suppressed by the unsteadiness of unconfined flow (Moss, 1972), or the high apparent viscosity of the bedload layer (Allen and Leeder, 1980), or sand being too coarse to form ripples (Harms et al., 1982), the scarcity of ripples suggests that the final flowing waters were ordinarily only a few centimeters deep.

A suspension blanket of mud overlying some coarse sand beds indicates that the sheetfloods occasionally scoured up enough silt and clay to develop a considerable suspended load, which settled out of the locally ponded waters left after the sheetfloods finished, as in the 1981 floods of South Africa described by Stear (1985). With drying of these standing bodies of water, the mudflat surface was sun-cracked (see the "shrinkage cracks" section of the flat-laminated sand sediment-type for interpretation).

Cross-laminated sub-type The solitary distribution of beds of the cross-laminated coarse sand sub-type indicates that they were deposited by the advance of an isolated, slipfaced bedform and not by trains of lower flow regime dunes, which deposit cosets of cross-beds

(Rubin and Hunter, 1982).

The lateral extent, internal continuity, and intercalation of cross-laminated coarse sands with desiccated sediments indicates that each cross-bed set was deposited during a single terrestrial sheetflood. As discussed in the cross-bedded sand sediment-type, any mound or depression in the path of a sheetflood could cause a slipface to develop because of the resulting flow seperation. Thus, a slipface could be generated where part of sheetflood's leading edge encountered sufficient topography on the mudflats. In their study of modern sheetflood deposits at Bijou Creek, Colorado, McKee, Crosby, and Berryhill (1967) found that low to high angle, planar-tabular cross-bed sets were commonly formed along the outer margins of depositional sheets "where water that had lost some of its force deposited sand along a sloping sediment front." Similarly, isolated, slipfaced bars could form within sheetfloods due to local rapid deposition as the flow spread out across the mudflats. Williams (1970) reports that flood deposits of ephemeral streams in central Australia include isolated sets of planar-tabular bar cross-beds which extend laterally for tens of meters.

PALEOCURRENTS

Due to their tight quartz cementation, ill-defined cross-beds, and scarcity, obtaining accurate measurements from cross-laminated coarse sands is extrememly difficult. Nonetheless, the attitudes of twelve sets of the cross-laminated coarse sand sub-type in a 600' interval of the upper Revett Formation at North Crow Creek were measured and analyzed as described in Appendix B: Paleocurrent Methodology. The pertinent data and results comprise table 2; figure 25 is the frequency distribution rose, for which n = 12, 0 = 271, and L = 33.6.

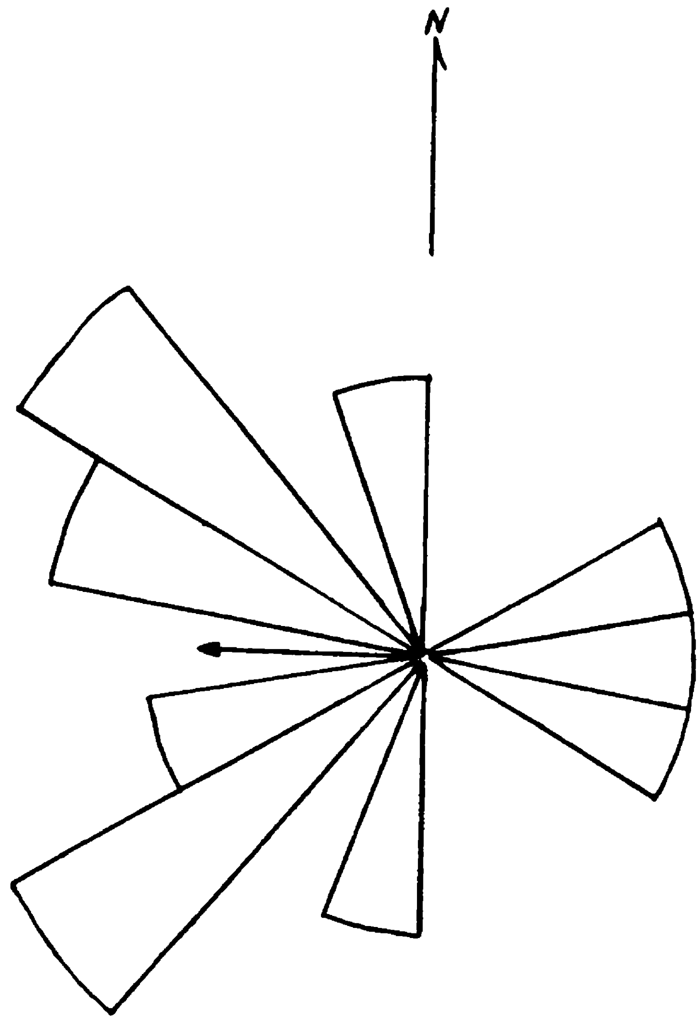
Although the set azimuth's in this analysis must differ somewhat from the mean flow direction of the currents that deposited them, taken as a group, they represent the dispersion about the mean flow direction (Smith, 1972). Thus, the vector mean azimuth of 271 degrees and the fact that nine of the twelve cross-bed sets represented here have a westward component suggest that the net flow and sediment transport directions were westerly.

The wide scatter (low consistency ratio, L) of set azimuths indicates that the currents depositing coarse sand cross-bed sets flowed in a broad spectrum of directions. This is as expected since ephemeral sheetfloods flowing out onto a flat surface spread out by the driving force of their hydraulic head, and therefore flow in all directions, even up gentle slopes. On nearly flat sufaces, the water surface slope determines the flow

FOOTAGE	Corrected Attitude	O	L	Set Thickness	Foresets	Occurrence
?	007/32se	097	100	4 cm	SA	S
?	042/44nw	312	100	3	C	S
?	025/42se	115	100	4	ST	S
?	342/40sw 278/21ne	310	85	6	SA	S
?	312/25sw	222	100	4	SA	S
?	036/40nw	306	100	5	C	S
?	345/34ne	075	100	10	ST	S
?	318/38sw	228	100	6	C	S
?	323/25sw	233	100	5	C	S
?	332/32sw	242	100	4	SA	S
?	018/34nw	288	100	3	SA	S
?	284/10sw	194	100	6	ST	S
?	026/40nw	296	100	3	C	S
?	086/12nw	356	100	7	C	S

Table 2: Paleocurrent data and statistics on sets of the cross-laminated coarse sand sub-type in the upper Revett Formation at North Crow Creek.

O: vector mean azimuth; L: consistency ratio
Footages: ? - footages unknown; data are listed by ascending order in the section.
Foresets: SA - straight angular; ST - straight tangential; C - concave.
Occurrence: S - solitary set.



n=12
O= 271
L = 33.6

Figure 25. Paleocurrent frequency distribution rose for the cross-bedded coarse sand sub-type in the upper Revett Formation at North Crow Creek. n = number of sets, O = mean vector azimuth (arrow on rose diagram), L = consistency ratio.

direction, to a greater extent than the ground slope (Mader and Teyssen, 1985; Alam et al., 1985). Thus, a slipface at the downstream edge of a sheetflood's depositional sheet, or a bar within a sheetflood will advance in any direction dictated by the flow's local expansion. Similarly widely dispersed flow patterns are reported from other ancient sheetflood deposits (Hubert and Hyde, 1982; Cheadle, 1986).

The extreme case of wide flow dispersal, set azimuths opposing the vector mean azimuth's direction (as in figure 25) often causes geologists to interpret the sediments involved as tidal deposits, regardless of their other qualities (McMechan, 1981; Mauk, 1983). However, recent advances in our understanding of the nature of flow in very low gradient fluvial systems (Alam et al., 1985; Mader and Teyssen, 1985) actually prescribe that local flow opposed to the net flow direction are to be expected in sheetflood systems. Indeed, bipolar cross-beds, and even herringbone cross-beds have been reported from ephemeral fluvial deposits (Mader and Teyssen, 1985; Alam et al., 1985), and Williams (1971) reports that bars in recent ephemeral flows locally had slipfaces facing upstream.

BEDDED SILT AND SAND SEDIMENT-TYPE

DESCRIPTION

The bedded silt and sand sediment-type comprises laterally continuous, sharply bounded, bedded sediments ranging from clayey silt, to silt, to sandy silt, to very fine grained sand. Most silt and sand beds thicken and thin slightly, but noticeably, across outcrops, although even beds are common. Because silty beds can be distinguished from sand beds in the field, their relative abundance is noted in the stratigraphic section of Appendix E.

Silty beds range from 1 to 15 centimeters thick; grain size generally decreases with bed thickness so that thinner beds originally contained more clay, and many thicker beds contain sand. Silty beds range from dark grey, to dark grey-green, to dark green. Some calcareous silty beds weather tan or light brown.

Beds of very fine-grained sand and silty-sand range from 5 to 30 centimeters thick. Most sand is well-sorted and white, but grey silty-sand is common.

Individual beds of the bedded silt and sand sediment-type can be traced across outcrops up to 10 meters wide, and none pinch-out on an outcrop scale. While beds of this sediment-type fine upward, in most cases the grading is very weak. Because all exposures of bedded silt and sand are thoroughly lichen-encrusted or limonite-stained due to the weathering of disseminated pyrite, determining the internal structure of beds is generally impossible. Internal laminations were discernable only in a few sawed samples, and in very rare, small, clean outcrop patches (figure 26) totaling less than five square feet in area in the upper interval of Unit B. Internal structure of stratigraphically lower silt and sand beds is unknown.

The few clean outcrop patches and sawed samples reveal featureless or weakly laminated silt and sand with occasional thinly-connected to unconnected, symmetrical lenses of coarse silt to very fine sand sharply overlain by laminations of finer sediment paralleling that undulating surface. The coarser lenses commonly occupy swales in the laminated layers (see figure 26). In two-dimensional views, silt lenses are 3 to 15 centimeters wide and less than 2 centimeters high. Though most show no internal structure, some lenses have laminations parallel to the upper or the lower bounding surfaces (figure 26). These sediments differ most clearly from lenticular couplets in the much greater width of their lenses, the fact that their lenses do not occur in trains, and in their lack of a graded muddy cap.

Meter-thick, fining- and thinning-upward sets of bedded silt overlain by sets of microlamina comprise the lower unit of the Prichard Formation at Blaine Mountain;

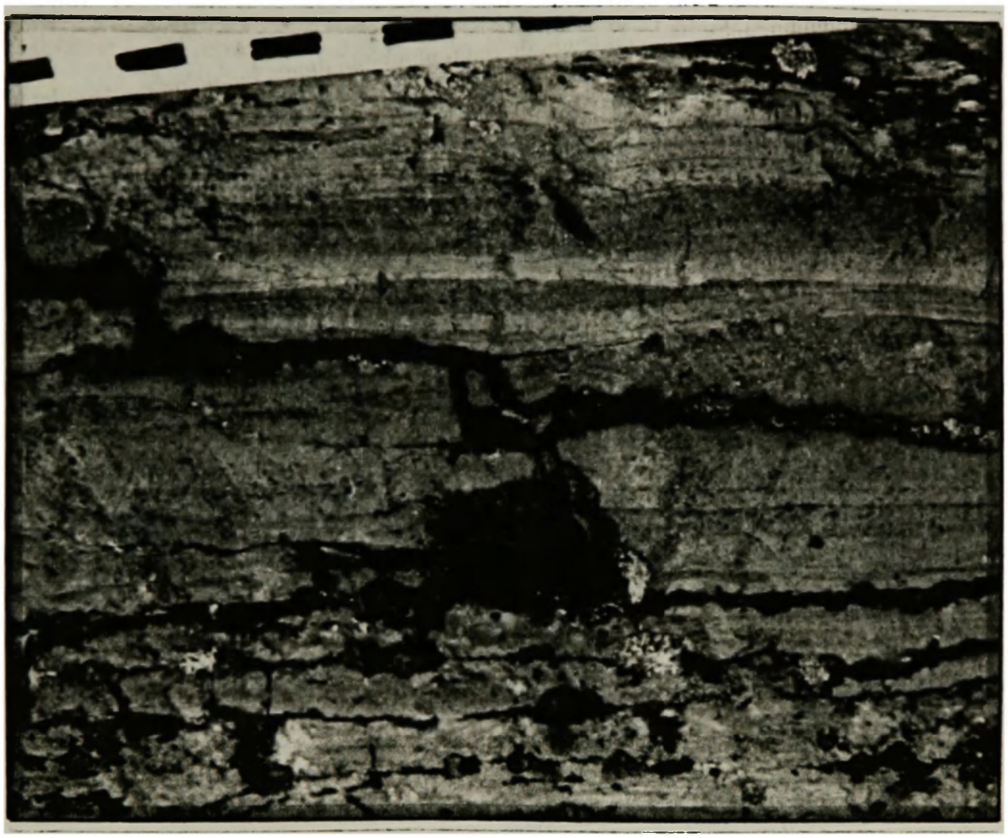


Figure 26. Bedded silt and sand sediment-type. Rare, clean outcrop showing internal structure of beds. Note coarser lenses (center and center right) occupying swales in finer, laminated layers.

the Prichard Formation's upper unit at Blaine Mountain, Lion Creek, and North Crow Creek consists of bedded silts and sands in no perceptible pattern, or rarely, in poorly-defined, fining-upward and coarsening-upward patterns. The bedded sands and silts intercalate, with sand beds becoming more abundant up-section.

INTERPRETATION

The lateral extent of silt and sand beds indicates that they were deposited from unconfined flows across broad flat surfaces, while their sharp contacts reveal that deposition occurred as a series of discrete events. The fine grain-sized comprising this sediment-type suggest that most of the sediment was transported in suspension. The grading of silt and sand beds records a steady decrease in the sediment transport during the depositional events, while the slight thinning and thickening of beds is probably due to loading, and suggests rapid sedimentation on a soft, saturated substrate. The absence of oscillation ripples confirms that sedimentation occurred below normal wave base in water on the order of several meters to tens of meters deep.

As the above inferences basically describe deposition from turbidity currents, the bulk of bedded silts and sands are interpreted as turbidites. Accordingly, turbid underflows scoured the sediment surface slightly and deposited fining-upward beds as the flow wained.

Because the lenticular sedimentary structures in some bedded silts and sands in the upper unit of the Prichard Formation are characterized by sporadic occurrence, occupation of swales, and mantling blanket of laminated finer sediment, they are interpreted as a form of hummocky stratification (Harms et al., 1984). Accordingly, large waves during storms scoured the bottom into oscillatory suspension and bedload transport. Decreased current strength or transport into deeper water (Allen, 1984, vol. II), deposited the bedload in swales, while oscillatory currents produced symmetrical bedforms. Finer sediment settled directly onto the undulatory surface or with brief traction sorting into laminations.

Many of the well-sorted, very fine-grained sand beds included in this sediment-type are similarly interpreted as storm deposits. Other clean sand beds probably represent sediments traction-transported by near shore currents. The increasing numbers of clean bedded sands up-section in the upper unit of the Prichard Formation records progressive shallowing.

MICROLAMINA SEDIMENT-TYPE

DESCRIPTION

The microlamina sediment type is characterized by even, sharply bounded, laterally continuous, millimeter-scale beds which grade from a tan, fine silt lower layer to a black, clayey upper layer. Microlaminae range from less than 0.25 millimeters up to about 3 millimeters thick and can be traced across outcrops a few meters wide.

Thin sections reveal narrow planes of homogenized sediment up to one centimeter long cutting across layering at high angles, and isolated grains of the medium- to coarse-grained sand population resting atop the clayey layers of microlaminae (figure 27).

The only occurrences of microlaminae are in the lower unit of the Prichard Formation at the base of the Mt. Blaine section (figure 28, Appendix E)), where fining- and thinning-upward, decimeters-thick sets cap thinning- and fining-upward, meter-thick sets of bedded silt (see the bedded silt and sand sediment-type).

INTERPRETATION

Microlaminae record suspensions of fine silt and clay settling out over broad areas. Silt grains deposited according to their settling velocities formed the lower layers of microlaminae. Eventually, the progressively depleted suspensions contained only clay particles, which deposited slowly to form the upper layers of microlaminae.

The absence of scours, ripple cross-laminations, and

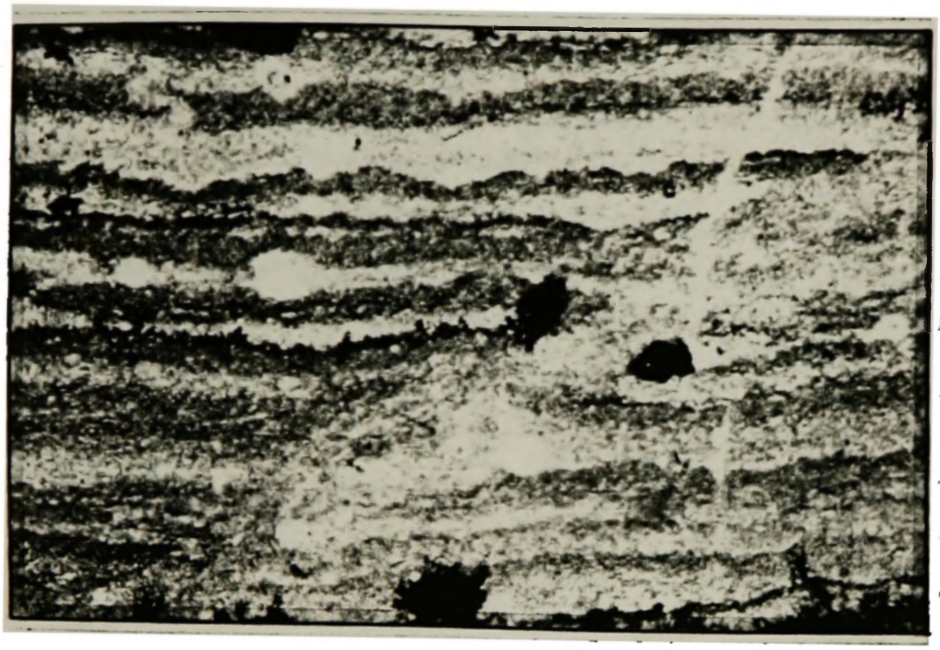


Figure 27. Microlamina sediment-type. Note high angle zone of homogenized sediment and scattered grains of medium sand.

flat-laminations confirms that microlamina were not deposited by traction transport; the absence of oscillation ripples indicates that the sediment surface was below fairweather wavebase.

Instead, microlaminae are interpreted as the most distal deposits of the turbidity currents which deposited beds of the bedded silt and sand sediment-type. The deposition of microlaminae via settle-out indicates that they record the lingering suspension which rolled outward from the terminus of active turbidity current transport.

The ruptures cross-cutting some microlaminae are interpreted as syn-sedimentary dikelets formed by settling, dewatering, or degassing.

Because the grains of medium- to coarse-grained sand scattered among microlaminae do not occur in the bedded silt turbidites and could not have been transported by the weak suspension clouds depositing microlamina, they were more probably blown from exposed sandflats to the east (Freeman and Winston, 1987) (see the coarse sand sediment-type). The fact that the sand grains rest atop microlaminae indicates that they were generally deposited in the time intervals between turbidity flows, rather than during the relatively brief periods the microlaminae represent.

CHAPTER 3

STRATIGRAPHY

This chapter describes the methods of litho-stratigraphic correlation, and provides an overview of the correlations and a unit by unit discussion of the stratigraphic units.

METHODS

Some stratigraphic intervals in the four measured sections in this study are dominated by a single sediment-type, while others are characterized by combinations of two sediment-types. These repeated intervals of single or paired sediment-types are here grouped into nine lithofacies which form widespread stratigraphic intervals and are therefore the basic elements of correlation between measured sections (figure 28, Appendix F in pocket). The nine lithofacies are: 1) cross-bedded and flat-laminated sand, 2) flat-laminated sand, 3) flat-laminated sand and even couples, 4) even couples and hematitic couplets, 5) hematitic couplets, 6) chloritic couplets, 7) coarse sand, 8) bedded silt and sand, and 9) microlamina and bedded silt. Lithofacies are correlated between measured sections on the basis of: 1) direct correlation of the same facies, and 2) gradual changes in facies (figure 28). Correlations based on gradual facies changes follow the interpretation developed in the chapter 2 on sediment-types that cross-bedded sands pass laterally to flat-laminated sands, which pass to even couples, which pass to couplets. The mudchip conglomerate sediment-type does not form continuous intervals that can be correlated from section to section.

Sets of alpha couplets and the distinctive bounding green sandy sediments form stratigraphic keybeds (see the couplet sediment-type) which are easily recognized in outcrop. Two keybeds occur in the North Crow Creek section, and one of these is correlated with the keybed in the Lion Creek section (figure 28).

OVERVIEW

Figure 28 and Appendix F illustrate the lithofacies correlations. Figure 29 shows previous formal assignment of the stratigraphic units, and figure 30 shows my proposed treatment of formal stratigraphic units based on the lithostratigraphic correlations.

The Burke, lower Revett, middle Revett, upper Revett, St. Regis, and Empire Formations (White and Winston, 1982; Alleman, 1983; Mauk, 1983; Harrison et. al, 1986) can be identified in the Mission Range (figures 28, 30). However, to the east in the Swan and Flathead Ranges, quartzite intervals of the Revett have not been mapped seperately and the entire Ravalli Group has been mapped as the Spokane or Grinnell Formation overlain by the Empire Formation (Winston, 1986b; Cronin et. al,

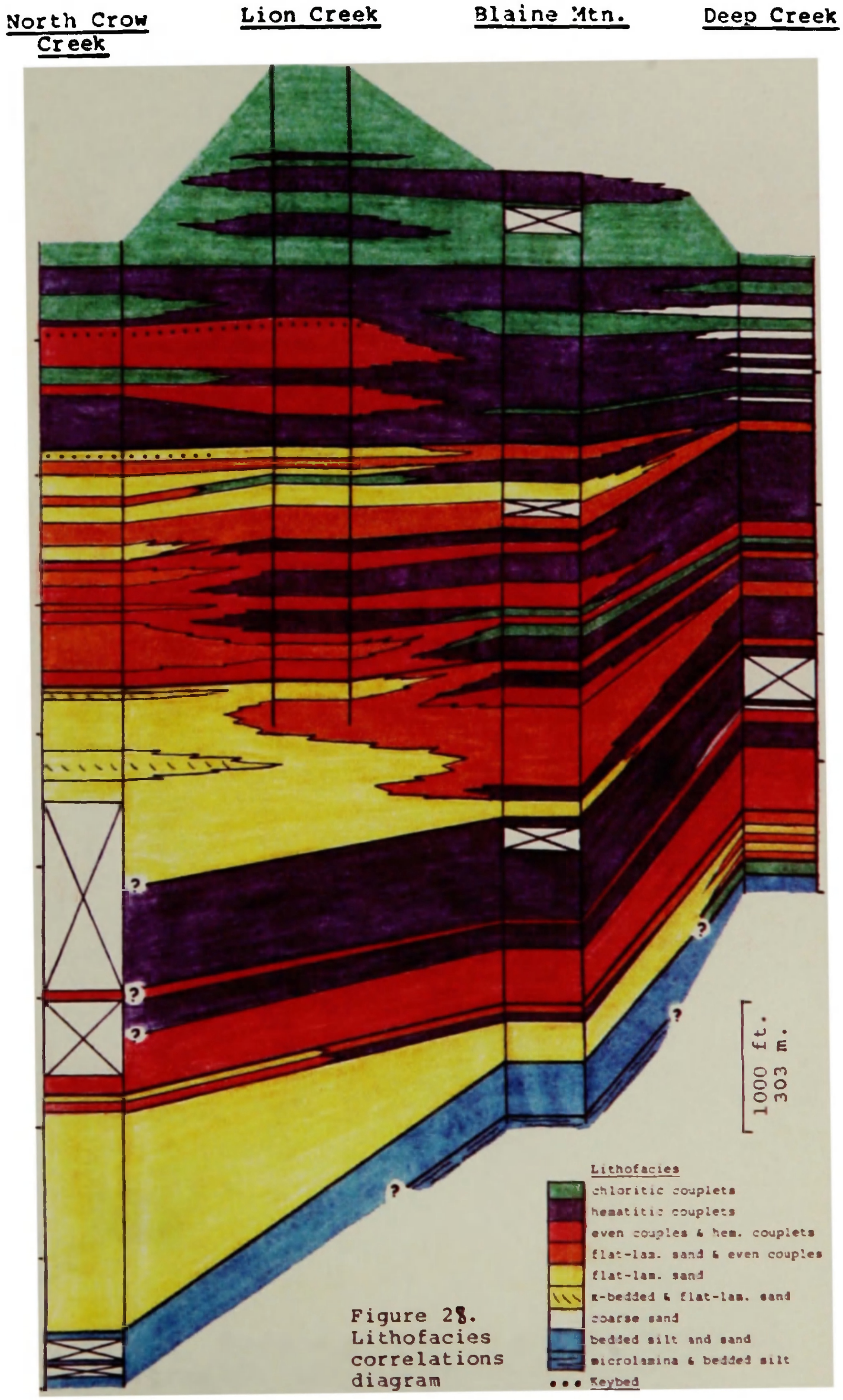


Figure 28. Lithofacies correlations diagram

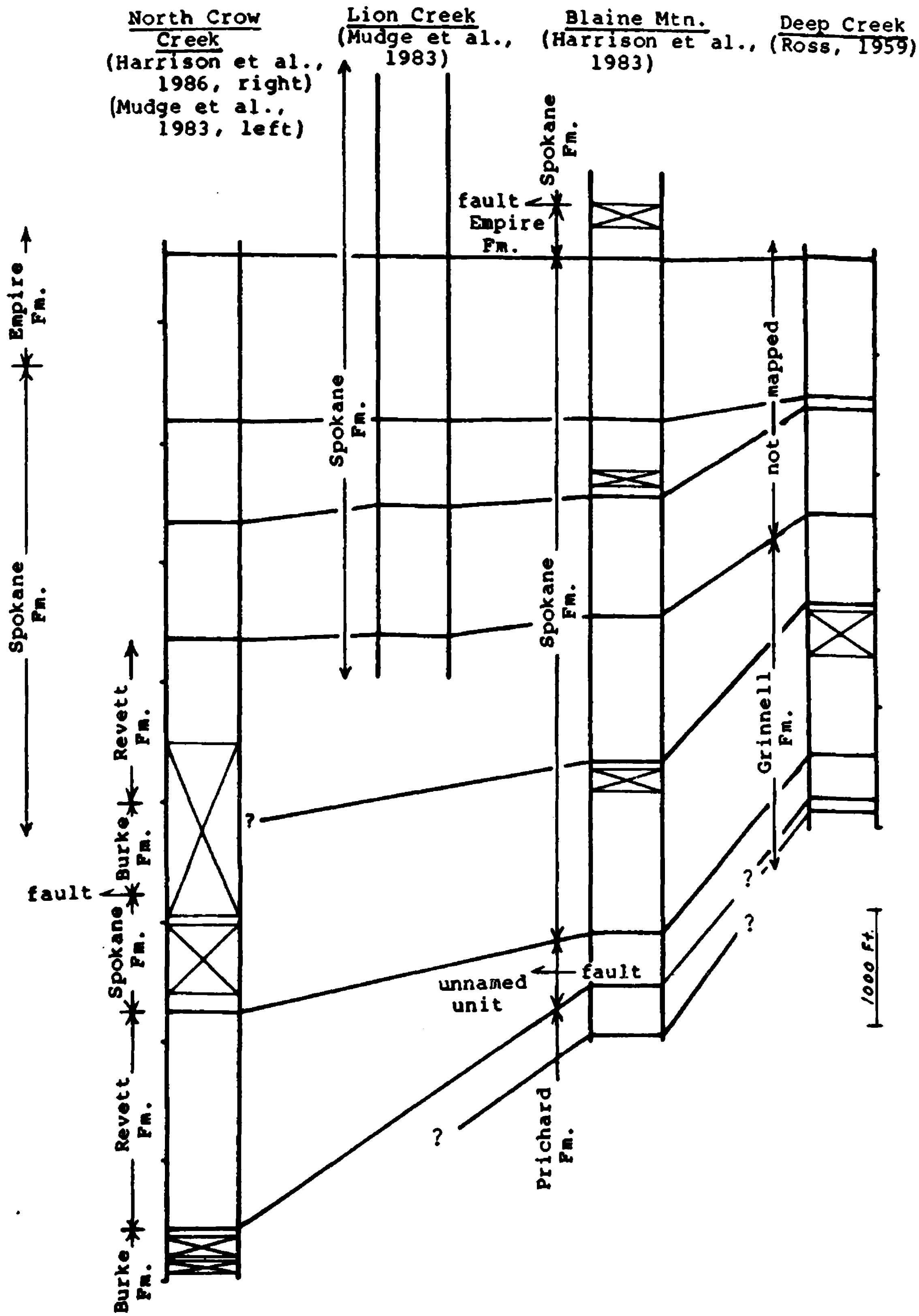


Figure 29. Previous authors designations of formal stratigraphic units plotted on the lithostratigraphic correlations of figure 28.

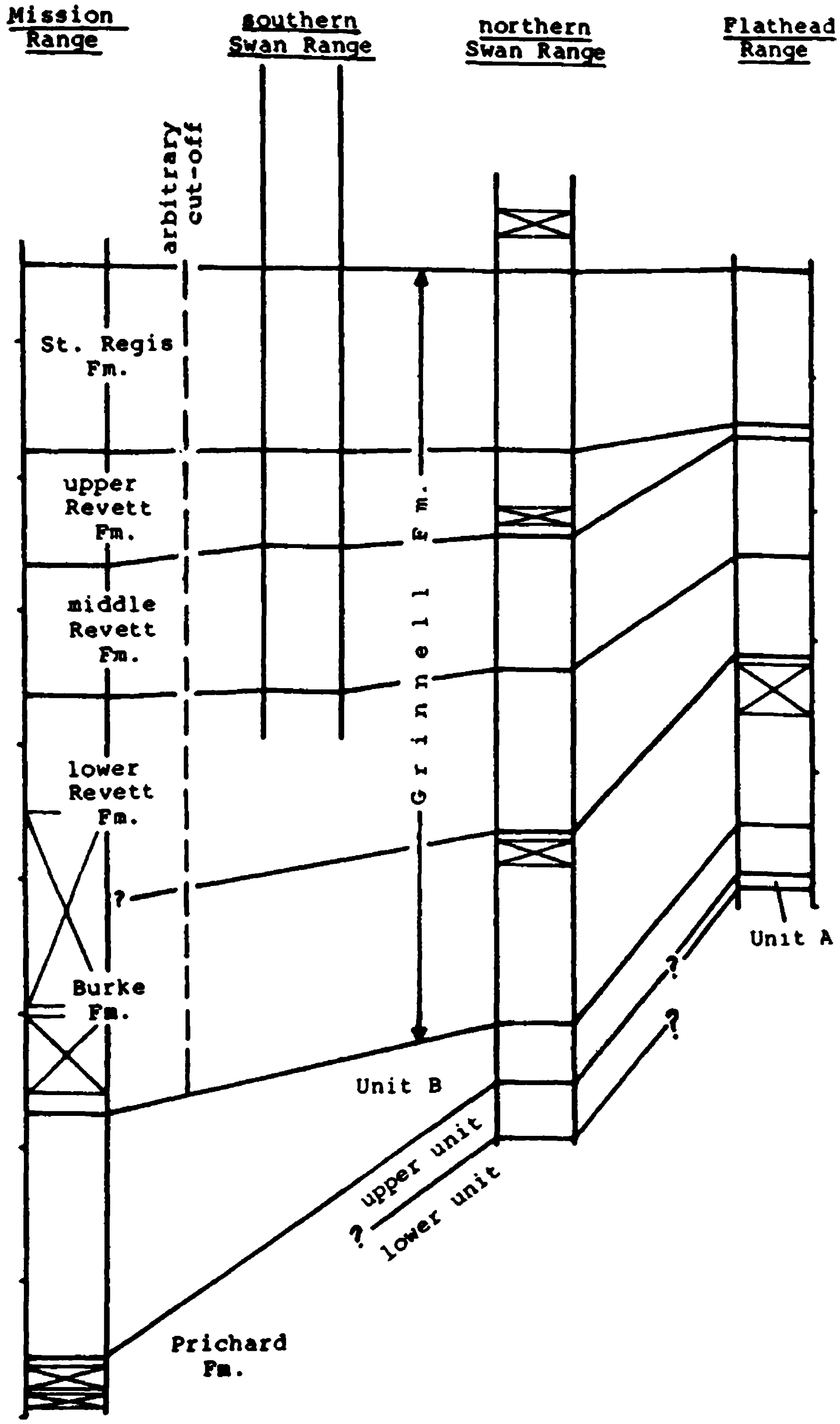


Figure 30. Proposed application of formal stratigraphy to the lithostratigraphic correlations of figure 28.

1986; Mudge et. al, 1983; Harrison et. al, 1983; Ross, 1959) (figures 29, 30). Thus, the Swan Valley between the Mission and Swan Ranges serves as an arbitrary cut-off (Wheeler and Mallory, 1953; North American Code of Stratigraphic Nomenclature, article 28c) between the western and eastern Ravalli Group nomenclatures. Because the entire Burke-Revett-St. Regis sequence correlates with the Grinnell (Spokane) Formation (figures 28, 30), not just the St. Regis Formation as has been proposed (Harrison, 1972) (figure 5), this thesis refers to "the Revett Formation and the correlative Grinnell Formation interval", for example.

The uppermost Prichard Formation was included in the measured sections because it bounds the lowermost Ravalli Group. However, the two units overlying the Prichard in the study area are informally designated Units A and B because their relationships with the formal stratigraphic units in the surrounding area are uncertain (figure 30).

PRICHARD FORMATION

The Prichard Formation has been mapped in the Swan Range (Harrison et al., 1983), but not in the Mission and Flathead Ranges. North of the study area, the Prichard has also been mapped in the Whitefish Range, and is tentatively recognized on the west side of Glacier National Park (Whipple et al., 1984).

The Prichard Formation in the Swan Range is here subdivided into informal lower and upper units at Blaine Mountain. These two units probably occur in the upper part of the 8,000 foot-thick Prichard (Ransome and Calkins, 1908; Hobbs et al., 1965). The upper unit also outcrops at North Crow Creek and Deep Creek. Correlation of these units with the surrounding area can not be attempted with the available data; the following descriptions are recorded as an aid to future, more detailed studies.

LOWER UNIT

Less than one hundred feet of intercalated microlamina and argillaceous silt beds at the base of the Blaine Mountain section comprise the lowest exposed unit of the Prichard Formation in the study area (figure 28). Scattered medium to coarse sand grains occur in thin sections of the microlamina.

This unit contains carbonate rhombs and pyrite cubes, and is rusty weathering. The contact between the lower unit and the upper unit is placed at the top of the highest set of microlaminae. This unit is mapped as Prichard Formation at Blaine Mountain (Harrison et al., 1983) (figure 29).

UPPER UNIT

The uppermost unit of the Prichard Formation in the study area consists of bedded siltstone and very fine-grained sandstone included in the bedded silt and sand sediment-type. In all three sections, argillaceous silty beds decrease upward, while clean, bedded very fine- to fine-grained sand increase greatly toward the top of the Prichard. At Blaine Mountain, the upper unit is 420 feet thick.

This unit is also characterized by dissenimated pyrite cubes that impart a rusty cast to the outcrops upon weathering; at Blaine Mountain and Deep Creek it includes abundant carbonate rhombs and their casts. Some brown-weathering, calcareous silty beds occur at Deep Creek.

The contact with the very fine-grained, flat-laminated sand beds of Unit B is gradational and is placed at the highest bedded silt characteristic of the Prichard Formation. Thus, the contact is placed by the same criterion used in the type area, the Coeur d'Alene District (Ransome and Calkins, 1908; Hobbs et al., 1965). This contact coincides with the lowest appearance of magnetite in the section. The upper unit's contact with Unit A at Deep Creek is covered.

Although the upper unit is mapped as Prichard Formation at Blaine Mountain (Harrison et al., 1983), at North Crow Creek, Harrison et al. (1986) mapped it as Burke Formation; and at Deep Creek, Ross (1959) mapped it as Grinnell Formation (figure 29).

Unit A

Unit A is a 100 foot-thick interval characterized by 1-3 centimeter-thick even and lenticular chloritic couplets, but including a few sets of hematitic even couplets and rare coarse sand beds. This unit is limited to the Deep Creek section where it occurs between the Prichard Formation and Unit B (figure 30). The lower contact of Unit A is covered and its uppercontact is placed at the base of the lowest flat-laminated sand bed of Unit B. Ross (1959) included this unit in the Grinnell Formation (figure 29).

To the east, Unit A may correlate with member 5 of the Appekunny Formation (Whipple et al., 1984), a bright green argillite unit characterized by thin couplets and microlaminae which forms the uppermost member of the Appekunny Formation on the east side of Glacier National Park, and separates the Prichard(?) Formation from the Grinnell Formation on the west side of Glacier National Park (Whipple et al., 1984). More detailed study of this unit is required to confirm this identification and correlation. The details of Unit A's pinchout between Deep Creek and Blaine Mountain are unknown.

Unit B

In its best exposure at North Crow Creek, Unit B consists of 1600 feet of flat-laminated sand overlain by 200 feet of flat-laminated sand, even couples, and hematitic couplets (figure 28). At Blaine Mountain, 300 feet of flat-laminated sand with a few sets of hematitic couplets is overlain by 120 feet of hematitic and chloritic couplets, even couples, and flat-laminated sand. In the Deep Creek section, 450 feet of silty flat-laminated sand, even couples and couplets includes occasional single beds and sets of coarse sand. In all three sections, sun-cracked mud drapes are common, and are especially abundant toward the top of the unit. Oscillation and current ripples on bedding planes occur throughout Unit A at North Crow Creek and Blaine Mountain, and at North Crow Creek the upper interval is particularly rich in sets of climbing ripples. In all sections, the upper contact of Unit B with the even couples and the couplets of the Burke Formation or the Grinnell Formation is gradational and placed at the top of the highest interval of flat-laminated sand beds.

At North Crow Creek Unit B is mapped as the Revett Formation (Harrison et al., 1986), but there is no doubt that Unit B in the Mission Range is not the Revett Formation repeated by faulting (figure 29), as indicated on the Wallace two-degree sheet (Harrison et. al, 1986). Although Unit B and the Revett Formation resemble each other superficially, Unit B is a single, thick interval composed almost entirely of flat-laminated sand, while the Revett comprises three members. The members include an abundance of flat-laminated sand, but they differ from

Unit B in that the lower Revett also contains abundant sets of the cross-bedded sand, even couple, and couplet sediment-types, and beds of the coarse sand sediment-type, and the middle and upper Revett also include copious sets of even couples, and couplets, and coarse sand beds (Appendix E or F).

At Blaine Mountain, Unit B is mapped as an unnamed formation (Harrison et al., 1983); cliff exposures of Unit B at the measured section confirm that it is not repeated by faulting as indicated on the preliminary Kalispell sheet (figure 29).

At Deep Creek, Unit B is included in the Grinnell Formation (Ross, 1959) (figure 29).

The status of this significant quartzite unit is uncertain, thus I designate it informally as Unit B.

In their study of the Coeur d'Alene District, Hobbs, Griggs, Wallace, and Campbell (1965) describe the transition zone between the Prichard and Burke Formations as a quartzite-dominated interval with mudcracks, ripple marks, and cross-stratification. They place the Prichard/Burke contact at the top of the highest laminated dark gray argillite interval characteristic of the Prichard; greenish grey argillite beds characteristic of the Burke occur among the predominant quartzite of the lower Burke Formation (Hobbs et al., 1965). This quartzite zone thins from a maximum of 2000 feet in the

Coeur d'Alene District, to 300 feet in the Libby Quadrangle, Montana (Gibson, 1948). Farther east into northwestern Montana, the zone does not include quartzites (Mauk, 1983; Harrison et al., 1986). Instead, the transition zone there consists of "Prichard-type laminated argillite interbedded with Burke-type green to grayish-green, thickly-laminated to thinly bedded magnetiferous siltite units with millimeter-scale argillite caps." (Mauk, 1983 p. 32); the lower part of the Burke consist of Burke-type siltites and argillites (Mauk, 1983).

Following the definition that the highest Prichard-type lithology marks the top of that formation (Ransome and Calkins, 1908; Hobbs et al., 1965; Mauk, 1983; Harrison et al., 1986), Unit B belongs in the Burke Formation, since, by definition it contains no Prichard-type lithologies and instead includes hematitic and chloritic argillite characteristic of the Burke. On the the other hand, assignment of Unit B to the Burke is inconsistent with that formation's character elsewhere in western Montana.

The southward thickening of Unit B from Blaine Mountain and Deep Creek, to the Mission Range, and its apparent absence to the immediate west (Mauk, 1983; Harrison et. al, 1986), suggests that sediments of this unit were derived from the south. Thus Unit B may be a distinct quartzite lobe at the same stratigraphic level as that in the Coeur d'Alene district (Ransome and Calkins, 1908; Hobbs et al., 1965) and the Salish Mountains of northwestern Montana (Loos, 1985), but spatially seperated from it. Additional detailed work on this stratigraphic level in and around the study area is required to define the formal stratigraphy of Unit B and resolve the complexities of the Prichard- Burke boundary.

BURKE FM.AND CORRELATIVE GRINNELL FM. INTERVAL

Most of the Burke Formation is covered at North Crow Creek. Only the basal 150 feet of even couples and hematitic couplets overlain by 90 feet of hematitic couplets, even couples, and flat-laminated sand beds are exposed (figure 28). The contact with, and lower portion of, the overlying lower Revett Formation are also covered. The thickness of this large covered interval was approximated by projecting the uniform dips of the lower Burke and lower Revett strata that were exposed (Appendix D).

Since bedding attitudes are consistent across the supposed fault, there is no evidence for a north-south normal fault through this covered interval as projected by Harrison and others (1986) (figure 30). That fault was mapped in the misidentification of Unit B as the Revett Formation; its inference was demanded to explain the apparent repetition of the Revett. Thus, although the 1500 feet of Burke at North Crow Creek is approximated from dips near the base and within the Burke (Appendix D), and by projecting the boundary of the Burke and lower Revett correlative intervals of the Grinnell Formation from Blaine Mountain (figure 28), its thickness is probably reliable within 100 to 150 feet.

The lower intervals of even couples and hematitic couplets exposed at North Crow Creek extend into the Blaine Mountain and Deep Creek sections, where the upper intervals of the Burke correlative interval of the Grinnell consists of hematitic couplets with rare coarse sand beds. A 25 foot-thick coarse sand interval occurs near the center of the Burke correlative interval of the Grinnell Formation at Deep Creek (figure 28).

At North Crow Creek, the Burke is mapped as the Spokane and Burke Formation juxtaposed by the inferred fault dismissed above (figure 30). At Blaine Mountain and Deep Creek, this interval is mapped as part of the Spokane (Harrison et al., 1983) and Grinnell (Ross, 1959) Formations.

REVETT FM. AND CORRELATIVE GRINNELL FM. INTERVAL

LOWER REVETT FM. AND CORRELATIVE GRINNELL FM. INTERVAL

The lower part of the lower Revett Formation is not exposed at North Crow Creek and the lower part of the correlative Grinnell interval is not exposed at Lion

Creek. However, correlating the sections at Blaine Mountain and Deep Creek with the lower Revett and correlative Grinnell intervals that do outcrop at North Crow Creek and Lion Creek permits extrapolation of the lower part of the lower Revett correlative Grinnell interval into the southern Swan Range and into the lower Revett of the Mission Range (figure 28).

The upper part of the lower Revett that does outcrop at North Crow Creek comprises two intervals of cross-bedded sand and flat-laminated sand separated by an interval of flat-laminated sand (figure 28 or Appendix F). The lower interval of cross-bedded and flat-laminated sand and the bounding intervals of flat-laminated sand pass northward to a thick interval of flat-laminated sand and even couples at Blaine Mountain (figure 28). The upper interval of cross-bedded and flat-laminated sand and the bounding intervals of flat-laminated sand at North Crow Creek pass to an interval of flat-laminated sand and even couples, and an interval of flat-laminated sand at Lion Creek, and to an interval of flat-laminated sand and even couples at Blaine Mountain (figure 28). The basal interval of flat-laminated sand, and the middle and upper intervals of flat-laminated sand and even couples at Blaine Mountain pass to three intervals of flat-laminated sand and even couples at Deep Creek. The coarse sand beds present in all four secitons of the lower Revett and correlative Grinnell become more abundant toward the northeast (Appendix F).

Alleman (1983) places the upper contact of the lower Revett "where vitreous white quartzite of the lower Revett passes upward to meters-thick intervals of thinly-laminated siltite-argillite rock-type separated by vitreous white quartzite beds centimeters thick." While there is no sediment-type analog for the siltite-argillite rock-type, that classification does include couplets. Thus, although Alleman's definition cannot be applied here exactly because of facies changes in the formation, his concept is preserved by placing the contact where flat-laminated sands of the lower Revett pass upward to meters-thick intervals of even couples and couplets. At North Crow Creek, the lowest occurrence of significant thicknesses of even couples and couplets is quite distinct (Appendix E or F).

The lower Revett Formation at North Crow Creek is mapped as the Revett Formation on the Wallace Quadrangle by Harrison et al. (1986), but is included in the Spokane Formation on the Choteau sheet (Mudge et al., 1983) (figure 30). In the Swan and Flathead Ranges, this unit is included in the Spokane (Harrison et al., 1986; Mudge et al., 1983) and Grinnell Formations (Ross, 1959).

MIDDLE REVETT FM. AND CORRELATIVE GRINNELL FM. INTERVAL

In the Mission Range, the middle Revett Formation comprises three intervals of flat-laminated sand and even couples, interstratified with intervals of even couples and hematitic couplets (figure 28). The lowest and thickest coarser interval carries eastward into the southern Swan Range, while the upper two coarser intervals pass to intervals of even couples and hematitic couplets. Northeastward into the northern Swan Range, all three coarser intervals pass to intervals of the even couples and hematitic couplets. In the Flathead Range, the lowest coarser unit is represented as an interval of even couples and hematitic couplets, while the upper two units pass to hematitic couplets. The intervals of even couples and hematitic couplets between the coarser units of the middle Revett Formation pass to hematitic and chloritic couplets to the east and northeast. The beds of coarse sand which occur in all four section of the middle Revett and correlative Grinnell Formation become increasingly abundant toward the northeast, and the top of the unit (Appendix F).

The contact of the middle Revett with the upper Revett at North Crow Creek coinncides with "the base of the lowest of a series of thick-bedded quartzite intervals above the finer-grained rock-types [sediment-types] of the middle Revett (Alleman, 1983 p.

39)" to the west. This contact is sharply marked by the lowest appearance of a 120 foot-thick unit of thick-bedded flat-laminated sand.

The middle Revett Formation at North Crow Creek is included in the interval mapped as the Spokane Formation (Mudge et al., 1983); at Lion Creek and Blaine Mountain, the unit is included in the Spokane Formation (Mudge et al., 1983; Harrison et al., 1986); and at Deep Creek it occurs in an area that was not mapped by Ross (1959) (figure 30).

UPPER REVETT FM. AND CORRELATIVE GRINNELL FM. INTERVAL

The upper Revett Formation in the North Crow Creek section comprises four intervals of flat-laminated sand separated by intervals of the flat-laminated sand and even couples. The three higher coarser units extend eaastward into the southern Swan Range. Northeastward, the upper two pass to an interval of flat-laminated sand and even couples, while one carries through the northern Swan Range and passes to the interval of flat-laminated sand and even couples which represents the upper Revett in the Flathead Range. All four sections include beds of coarse sand, which become more abundant to the northeast. The Blaine Mountain section is especially rich in coarse sand beds and sets, and includes sub-type 1 mudchip conglomerate beds in its coupleted intervals. A stratigraphic keybed (the couplet sediment-type) occurs high in the Mission Range section (figure 28).

Mauk's (1983) placement of the upper Revett-St. Regis contact at the top of "the highest thick interval of thick-bedded quartzite." cannot be followed precisely because, due to facies changes, the upper quartzitic member at North Crow Creek is dominated by thin-bedded quartzites. However, his concept of the upper Revett is preserved by placing the contact at the top of the highest thin-bedded quartzite interval. At North Crow Creek this contact is clearly defined, as the overlying St. Regis Formation is devoid of the fine-grained quartzite beds which dominate the upper Revett.

The upper Revett Formation at North Crow Creek is included in the interval mapped as Spokane Formation (Mudge et al, 1983). At Lion Creek Mudge et. al, 1983) and Blaine Mountain (Harrison et. al, 1986) the upper Revett correlative interval of the Grinnell Formaion is included in the Spokane Formation, and at Deep Creek it occurs in an area that was not mapped by Ross (1959) (figure 30).

ST. REGIS FM AND CORRELATIVE GRINNELL FM. INTERVAL

The name St. Regis is applied to the redbeds overlying the Revett Formation at North Crow Creek even though they are included in the interval mapped as the Spokane Formation (Mudge et. al, 1983) (figure 29) because the lithofacies correlations (figure 28) demonstrate that the Grinnell (Spokane) Formation is the eastern facies of the Burke-Revett-St. Regis sequence, and not a separate clastic wedge that correlates solely with the St. Regis (Harrison, 1972; Harrison et. al, 1986). Since the Revett, which allows tripartite division of the Ravalli Group, has been identified in the Mission Range (Harrison et. al, 1986; this study), it is appropriate that all three western names be employed.

The St. Regis Formation in the Mission Range comprises intervals of chloritic couplets, hematitic couplets, and even couples and hematitic couplets (figure 28). The two even couple-bearing intervals extend into the southern Swan Range. A stratigraphic keybed (see the couplet sediment-type) in the upper couple-bearing interval also extends from North Crow Creek into Lion Creek (figure 28). In the northern Swan Range, the St. Regis correlative interval of the Grinnell Formation consists of hematitic and chloritic couplets. Both Swan Range sections include locally abundant sub-type 1 mudchip conglomerates (Appendix E). Coarse sand beds occur in all four section (Appendix F), but become strikingly more abundant to the northeast, until at Deep Creek, meters-thick sets of coarse sand comprise nearly half of the unit's thickness.

As implied by Harrison's (1972, p. 1222) discussion of the green beds below the Middle Belt Carbonate, no widely accepted lithologic criteria for defining the upper contact of the St. Regis and Spokane Formations exists. "Traditionally, the contact between the Spokane [and St. Regis] Formation[s] and the Empire Formation is placed at the boundary where the dominating color of the section changes from red to green." (Whipple, 1980). Despite the inconsistencies in mapping and correlation that result from using this diagenetically-produced parameter, the color change remains the principal criterion for drawing that contact (Whipple, 1980; Winston, 1987a).

The dilemna of this contact is illustrated by three recent studies in the eastern Belt basin. In a detailed study of the Spokane-Empire transition in the Rocky Mountain Disturbed Belt of west-central Montana, Whipple (1980) found that a transition zone of alternating red- and green-bed sequences bounds the two formations. He placed this zone in the Spokane Formation, thus drawing the contact at the top of the highest thick redbed sequence. However, on the Choteau quadrangle and particularly at North Crow Creek, "purple beds of predominantly argillite occur in the lower part of the [Empire] formation (Mudge et al, 1983).", and on the Wallace sheet (Harrison et al., 1986), red argillite intervals occur near the middle of the Empire Formation.

In this thesis, the St. Regis boundary is placed at the top of the highest thick interval of hematitic couplets, above which the dominant color of the section is green. The resulting, remarkably consistent thickness of the St. Regis Formation and correlative Grinnell Formation across the study area suggests that, at least in the three ranges involved, the color change at this boundary is closely related to original stratification units

The redbeds in the lower third of the St. Regis Formation at North Crow Creek were mapped as the Spokane Formation by Mudge et al. (1983), while the alternating hematitic and chloritic units forming the upper two-thirds of the St. Regis were included in the Empire Formation by Mudge et al. (1983) (figure 30). At Lion Creek, Mudge et al., (1983) mapped the Spokane-Empire boundary more than 1500 stratigraphic feet higher than the boundary assigned here; at Blaine Mountain, the boundary as mapped by Harrison et al. (1983) coincides with that used here; and at Deep Creek, it occurs in an area not mapped by Ross (1959).

EMPIRE FORMATION

In all sections, the Empire Formation comprises the green and calcareous green argillite-dominated unit above the redbed-dominated St. Regis Formation and correlative Grinnell Formation interval (figure 28 and Appendix F). All sections of the Empire include coarse sand beds.

The lowermost 200 feet of the Empire Formation measured at the North Crow Creek section includes recessive-weathering calcite nodules characteristic of that formation (Harrison et al., 1986), and are overlain by at least 500 feet of unmeasured calcareous green-beds which include locally abundant calcite nodules and pyrite cubes up to 3 centimeters across.

At Lion Creek, green and calcareous green argillite dominates the Empire Formation, although a 180 foot-thick redbed sequence occurs 200 feet above the base of the formation (figure 28). The couplets constituting most of that sequence are distinctly thinner (generally < 1 centimeter) and muddier than those in the underlying Spokane Formation, and the interbedded sub-type 2 mudchip conglomerates are unique to the Empire (Appendix E). The redbed intervals which occur higher in the formation include copious calcareous green argillite sequences (Appendix E or F), and the uppermost 150 feet of the Lion Creek section includes recessive-weathering calcareous nodules characteristic of the Empire (Harrison et al., 1986). Possibly, a red bed sequence even higher in that formation led Mudge et al. (1983) to draw the Spokane/Empire contact within the Empire Formation (figure 30).

The Empire/Spokane contact mapped at Blaine Mountain by Harrison et al. (1983) coincides with my placement.

However, the down-to-the-west fault mapped by Harrison et al. (1983) repeating the upper Spokane above the lower Empire, apparently to account for the redbeds above the green beds of the lower Empire was not evident where the section was measured, as bedding dipped uniformly across the inferred fault. In addition, the distinctive sub-type 3 mudchip conglomerate beds that characterize the supposedly fault-repeated redbeds do not occur in the Spokane and thus demonstrate that the uppermost redbeds are not fault-repeatd Spokane, but are a distinct hematitic unit low in the Empire. Correlation of this interval with an interval of alternating red- and green-beds within the Empire Formation at Lion Creek supports this conclusion (figure 28).

The 150 feet of the lowermost Empire Formation measured at the top of the Deep Creek section comprises sequences of green and calcareous green argillite intercalated with decimeters- to meters-thick sets of coarse sand. Green-beds continued upward in cliffs for at least 250 foot-thick above the top of the measured section.

Chapter 4

SYNTHESIS

SEDIMENTOLOGIC INTERPRETATION

Winston (1986c) developed the sediment-type methodology used in this study. He also proposed a depositional model for the western facies of the Ravalli Group and speculated about sedimentation on the eastern side of the Belt basin. In this section, Winston's (1986c) basinwide depositional model is reviewed, followed by a depositional model for the eastern side of the basin based on data and interpretations of this study. Finally, my depositional model is compared with Winston's

WINSTON'S DEPOSITIONAL MODEL

Winston (1986c) demonstrated that his cross-bedded sand sediment-type passed downslope (northward and eastward) to the flat-laminated sand sediment-type and speculated that in the Ravalli Group these passed basinward to the even couple and discontinuous layer sediment-types, and that both of these passed basinward to the even couplet and lenticular couplet sediment-types. He observed that the Ravalli Group's medium- to coarse-grained sand population was concentrated along the northeastern side of the basin, and speculated that these sands occurred dominantly in the form of the locally oolitic, low-angle cross-bedded lenses and beds comprising his coarse sand and intraclast sediment-type. Winston (1986c p. 106) interpreted this basinward progression of sediment-types in the following manner:

"Floods that drained uplifted blocks south and west of the basin carried medium to fine sand and mud northward and eastward (Mauk, 1983) into the basin through channels 1 to 2 meters (3 to 6 ft.) deep that traversed the middle parts of large alluvial aprons. Flow in these channels was in both the upper part of the lower flow regime, forming planar crossbeds, and the upper regime, forming epsilon crossbeds in which laminae conform to the channel outlines. During floods, water may have spilled out onto lateral flood plains, depositing fine sand, silt and clay of the discontinuous layer sediment type. As channels descended across the alluvial aprons, slope and flow velocity diminished, so that only fine sand and mud were transported. Flow depth decreased, and flow shifted to the upper regime, forming sheetfloods that deposited beds of the flat-laminated sand sediment type on the apron toe and deposited the lower sandy layers of the even couple sediment type out on the sandflats. Flow continued out onto exposed mudflats, ripping up dried mud polygons and depositing fine sand and silt half-couplets of the even couplet sediment type. Where waters ponded, clay settled from suspension, forming clay caps. As sheetfloods on the alluvial aprons began to wane, velocity diminished, so that flow shifted to the lower part of the lower flow regime, depositing rippled silt and clay of the discontinuous layer sediment type. Flow ceased, ponds dried, and the clay soon desiccated and cracked. Where water lingered for longer periods, silt and clay were reworked into ripples of the lenticular couplet sediment type, and beaches and shoals of the coarse sand and intraclast sediment type were built, mostly along the eastern margin of the basin. Distal surfaces that were continuously submerged occasionally received thin layers of the microlamina sediment-type." (Winston, 1986c).

PROPOSED DEPOSITIONAL MODEL

The lithofacies correlations (figure 28) and vertical sequences in the measured sections (Appendix E) indicate the following basinward progression of the dominant Ravalli Group sediment-types across the study area: 1) cross-bedded sand interbedded with flat-laminated sand, 2) flat-laminated sand, 3) even couples, and 4) even and lenticular couplets with interbeds of coarse sand, and less commonly, sub-type 1 mudchip conglomerate. Integrating the depositional processes of these sediment-types with their basinward progression on the scale of a single flood-event produces a single, hypothetical flood deposit (figure 31), while integrating them on a geographical scale produces a model

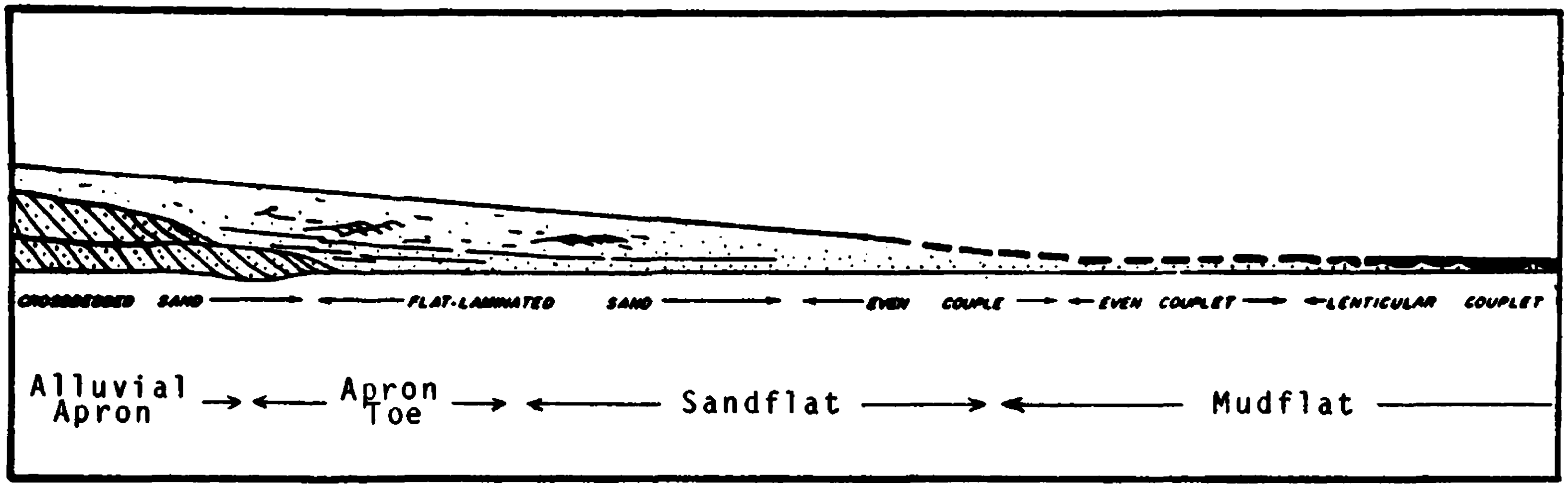
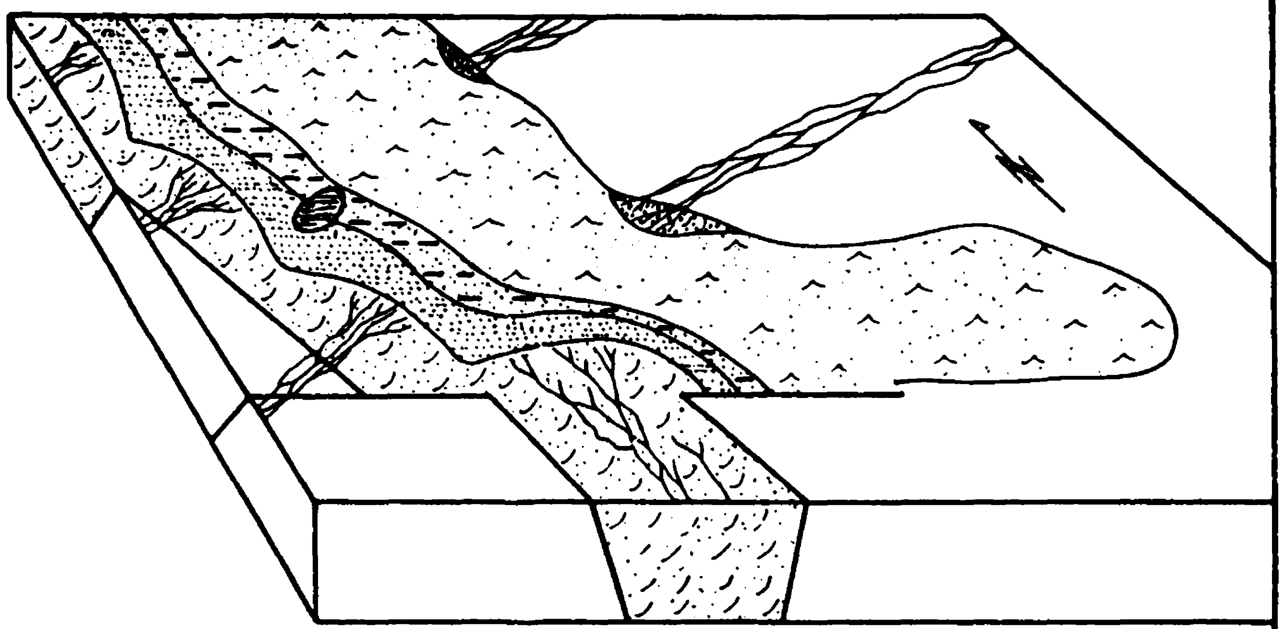


Figure 31. Diagramatic sediment-type cross-section of a single, hypothetical flood deposit in the Ravalli Group from low on the alluvial apron, down across the playa mudflats (modified from Winston, 1986c). (Also see figure 32 text.)

Figure 32. Hypothetical block diagram of the Ravalli Group depositional environment and sediment-type facies tract: interbedded cross-bedded sand and flat-laminated sand sediment-types, deposited by braided streams and proximal sheetfloods low on the alluvial aprons characterize the western and southern parts of the basin. This facies passes basinward onto the toe of the alluvial apron and the sandflats to the flat-laminated sand sediment-type deposited by sheetfloods. Flood deposits thinned basinward to the even couple sediment-type, which accumulated from the sandflat, out onto the margins of the playa mudflats. Across the mudflats, individual flood deposits continued to thin and became even couplets. Submerged surfaces were locally reworked by waves into the lenticular couplet sediment-type. Rare perennial lakes on the sandflats were muddied by sheetfloods flowing in from the south and southwest, and silt and clay settled from suspension to form alpha couplets. Occasional sheetfloods from the east flowed down localized alluvial aprons and spread out across the exposed mudflats, depositing medium- to coarse-grained sand derived from the craton as beds of the coarse sand sediment-type (text and diagram modified from Winston, 1986c).



CHARACTERISTIC SEDIMENT-TYPES	DEPOSITIONAL SUB-ENVIRONMENT
Cross-bedded sand	Alluvial Aprons
Flat-laminated sand	Apron Toe and Sandflats
Even couple	Sandflats
Even and lenticular couplet	Mudflats
Alpha couplet	Perennial Lake
Coarse sand	Alluvial Aprons and Mudflats

Figure 32.

of the Ravalli Group depositional environment (figure 32).

Accordingly, braided rivers flowing down vast alluvial aprons that sloped into the study area from the south and southwest (Bowden, 1977; Mauk, 1983; Alleman, 1983; Greene, 1984; Winston, 1986c) (figure 32) became progressively less channelized and passed downstream to braided flow within sheetfloods as the apron's slope decreased. Bars of fine- to very fine-grained sand migrating in the braided flow within the proximal sheetfloods deposited sets of the cross-bedded sand sediment-type interbedded with beds of the flat-laminated sand sediment-type (figure 31). Flow velocity and flow depth decreased as the sheetfloods spread out across the toe of the alluvial apron and out onto sandflats. Well-sorted, fine- to very-fine grained sand transported as bedload was deposited in flat-laminations as the flow strength diminshed and the upper plane bed aggraded, forming clean beds of the flat-laminated sand sediment-type. Continued deceleration and thinning of the flow caused its capacity for suspended sediment to progressively diminish, so very fine sand, silt, and clay settled toward the upper plane bed from the overloaded supension. These fine sediments underwent brief traction transport, and were deposited together as the bed rapidly aggraded, forming the silty-sand beds of the flat-laminated sand sediment-type, and the lower sandy-silt layers of the even couple sediment-type (figures 31, 32). Although in dropping the excess suspended load, all grain-sizes present were included in the deposit, the higher settling rate of the coarser grains caused the suspended load to fine progressively as it traveled. Because of the basinward depletion of coarser grains, a distally-fining flood deposit was formed.

The ever-expanding, ever-thinning, muddy, upper regime sheetfloods flowed across the sandflats and out onto exposed mudflats. The vast sheet of flowing water ripped-up dried mud polygons littering the surface of the mudflats and transported them as bedload. The continuing deceleration of the flow forced the mudchips and the silt and clay settling into the bedload from the overloaded suspension to be deposited, thus forming the lower silty layers of even couplets (figures 31, 32). Occasional sheetfloods on the distal mudflats scoured-up enough mudchips to attain the consistency of cohesive debris flows, and simply congealed upon deceleration to form some beds of the sub-type 1 mudchip conglomerate sediment-type

Waters ponded locally on the sandflats and over vast tracts of the mudflats as each flood wained. Where standing water lingered for long periods, silt and clay were reworked by waves into the symmetrical ripples which cap some even couples and flat-laminated sand beds, and which comprise the lower layers of lenticular couplets. Clay settled from muddy standing water directly onto the bed, forming the clay layers capping lenticular couplets, even couplets, even couples, and many flat-laminated sand beds. As the ponds dried-up, the clay layers desiccated and cracked. The repeated development of different crack patterns on alternately saturated and desiccated, slowly aggrading parts of the distal mudflats brecciated the upper few centimeters of the mudflats and formed other beds of the sub-type 1 mudchip conglomerate.

Rarely, large perennial lakes fed by springs occupied broad shallow depressions in the distal sandflats or proximal mudflats. Sheetfloods flowing into the lakes muddied their waters, and silt and clay settling from the suspension formed alpha couplets (figure 32).

Upper regime sheetfloods from the east carrying medium- to coarse-grained quartz sand flowed down local alluvial aprons and spread out across the exposed mudflats. The floods scoured up mud polygons from the mudflats and transported them with the coarse sand bedload. As the sheetfloods thinned and decelerated, the mudchips and coarse sand were deposited on the aggrading upper plane bed, forming beds of the bedded coarse sand sub-type. Advancing slip-faces at the leading edge of the depositional sheets, and isolated bars within the sheetfloods deposited sets of the cross-bedded coarse sand sub-type. Occasional eastern sheetfloods inundated and crossed the mudflats, reaching the sandflats, or even the apron toe on the western side of the basin, so that coarse sand beds were interbedded with beds of the even couple or flat-laminated sand sediment-types.

That sheetfloods flowing in from the south, southwest, and east all crossed the same mudflats confirms that the Ravalli Group landscape was a vast plain, and requires that standing water did not impede the flow of sheetfloods across that plain. Apparently, the water contained in the sheetfloods spread out over such a large area that the playa lakes which formed, quickly evaporated, or were absorbed and became subsurface flow which left the basin. Many sheetfloods probably continued spreading out across the flats until they simply thinned and vanished. During Ravalli Group time then, what has been called the Belt "sea" (Winston, 1986c) was probably only a series of geographically broad, but only centimeters-deep, ephemeral playas on the vast mudflats. Since the term sea gives the impression that a single, deep, perennial water body existed during the deposition of the Ravalli Group, the scattered, ephemeral playas envisaged here, taken together, are instead referred to as the ephemeral Belt lake.

Interestingly, all the stratigraphic units thin slightly to the north and east (figure 30). Moreover, the basinward thinning of their fine sand, silt, and clay component is actually even greater, since the abundance of coarse sand beds decreases to the west. Thus, the alluvial apron-sandflats-mudflats complexes that derived from the south and southwest thin and fine in the distal direction, as is characteristic of alluvial systems. And, the slightness of the thinning, relative to the lateral distances involved (tens of miles), confirms that the slope of the depositional surface was astonishingly low, and that the Ravalli Group depositional environment was indeed a vast plain.

COMPARISON

In both the model proposed here and the one proposed by Winston (1986c), braided rivers traversing large alluvial aprons pass downstream to sheetfloods which spread out across vast sandflats and mudflats, and left ponded waters on the flats after the floods ended. However, although the two interpretations are in general agreement, there are some important differences in their details, and several features of the depositional system are newly recognized by this study. The points of contrast and the new concepts are enumerated and discussed in the order of their position in the proximal-to-distal depositonal progression described above.

1) Because Winston's studies of the Revett Formation (White and Winston, 1982; White, Winston, and Mauk, 1984; Winston, 1986c) concentrated on its more proximal western facies, his cross-bedded sand sediment-type comprised the epsilon cross-beds formed by lateral channel migration, plus the megaripple and bar cross-beds deposited within those channels. In contrast, the cross-bedded sand sediment-type of this thesis is interpreted to record the migration of bars where channellized braided flow graded to sheetfloods, or what has been referred to herein as braided flow within proximal sheetfloods.

2) This study includes the new interpretation that suspension-with-traction sedimentation characterized deposition from the southern and southwestern sheetfloods. Accordingly, the continuous deceleration of the sheetfloods decreased their capacity for suspended sediment so that fine sand, silt, and clay settled toward the bed from the overloaded suspensions and briefly became part of the bedload, depositing the flat-laminated sand sediment-type and the lower layers of the even couple and even couplet sediment-types. The preferential deposition of the coarser grains upslope produced a distally fining load and deposit. Thus, theoretically, a

flat-laminated sand bed passes basinward to the lower fine sandy-silt layer of an even couple, which passes to the lower, thin, silty layer of an even couplet.

3) Winston's interpretation of the discontinuous, fine sand and silt ripple cross-beds, and vaguely defined mud lenses that characterize his discontinuous layer sediment-type as distal sheetflood deposits is not borne out, because the progression of flat-laminated sand to even couples continues basinward to couplets without passing through a tract of the discontinuous layer sediment-type. Since the discontinuous layer sediment-type has proven to be a proximal sedimentary species, Winston's other interpretations of it as waining stage or overbank deposits of the braided streams are more probably correct.

4) Mudchip conglomerate beds are a newly recognized sedimentary taxon in the Ravalli Group. Sub-type 1 mudchip conglomerates are interpreted as cohesive debris flows or as desiccation breccias; sub-type 2 conglomerates are interpreted as sheetflood deposits; and sub-type 3 conglomerates are interpreted as lacustrine beaches or shoals washed by waves.

5) The most important difference between the two environmental interpretations compared here concerns the ponded waters left on the mudflats after floods, and the significance of the Ravalli Group's coarse sand beds.

Winston (1986c) speculates that an essentially permanent lake or sea regularly expanded and retreated across the Ravalli Group mudflats. Coarse sands from a thick blanket on the crystalline basement north of the Helena Embayment (Winston, 1986c; Frost and Winston, 1987; Freeman and Winston, 1987) was transported around the basin by longshore drift and worked into small beaches and shoals along the temporary shorelines. In contrast, this study concludes that the Belt lake usually dried up quickly and completely after floods, and the playa floor was actually dry most of the time. Sheetfloods from the craton carried coarse sand across the exposed mudflats, and even as far west as the sandflats and apron toe.

6) In addition to the vast, playa lakes that periodically covered the mudflats, this study proposes that the Ravalli Group landscape rarely included the true perennial lakes in which alpha couplets accumulated.

7) Winston (1986c) interprets apparent bipolar cross-beds of his coarse sand and intraclast sediment-type as beach foresets and the accompanying washover fan cross-beds. In contrast, bipolarity in the paleocurrent roses of both the coarse sand and cross-bedded sand sediment-types of this study is interpreted primarily as a result of the Ravalli Group's nearly level gradient. That is, sheetfloods flowing out onto the flats are believed to have been the highest

topographic features on the Ravalli Group plain, and to have thus spread out in many directions by the driving force of their hydraulic head.

DEPOSITIONAL HISTORY

The depositional history of Unit A, Unit B, and the Ravalli Group of the study area is interpreted by synthesizing the depositional model (figures 31, 32) and the lithofacies correlations (figure 28).

The coarsening-upward sequence recorded by the upward progression from the microlamina sediment-type to the bedded silt and sand sediment-type in the upper units of the Prichard Formation record shallowing upward interpreted by Loos (1985) and Finch and Baldwin (1984) as a prograding turbidite fan complex that prograded down the axis of the Belt basin, and filled the deep, perennial Belt lake. The northward- and eastward-fining and -thinning beds of the flat-laminated sand, even couple, and couplet sediment-types of Unit B record terrestrial sheetfloods that flowed from the south, transporting fine sand, silt, and clay, and forming an alluvial apron-sandflat-mudflat complex. Occasional sheetfloods from the east transported coarse sand from the craton, ripped up mud polygons, and deposited beds of the coarse sand sediment-type as they flowed across the desiccated mudflats (Appendix F). Whether the even and lenticular couplets of Unit A are a mudflat deposit at the distal end of this alluvial system, or a shallow platform deposit marginal to the Prichard turbidite fan complex should be clearer with additional study. As the alluvial apron and sandflats of Unit B retreated, the mudflats expanded to the south and southwest, depositing the even couples, and even and lenticular couplets of the Burke Formation and the correlative interval of the Grinnell Formation. The advance of a small alluvial apron from the east deposited a thick set of coarse sand beds near the middle of the Burke correlative interval of the Grinnell Formation at Deep Creek (figure 28).

A major advance of alluvial aprons from the southwest is recorded by the lower Revett Formation. Two intervals composed of sets the cross-bedded sand sediment-type interbedded with beds of the flat-laminated sand sediment-type record the farthest advance of shallow braided streams down the alluvial aprons. The lower Revett apron sloped northeastward into the depocenter along the eastern part of the basin, passing to the sandflats and alternately submerged and exposed mudflats recorded by beds of the flat-laminated sand, even couple, and couplet sediment-types of the correlative interval of the Grinnell Formation. Occasional sheetfloods from the east deposited coarse sand beds on the mudflats, and even as far west as the sandflats and toes of the lower Revett alluvial apron (Appendix F).

The mudflats expanded southwestward over the sandflats and alluvial apron, although three minor advances of the sandflats deposited the wedges of flat-laminated sand and even couples in the middle Revett Formation. The alluvial pulses are reflected in the correlative interval of the Grinnell Formation as three coarser intervals of even couples and hematitic couplets intercalated with intervals of even and lenticular couplets (figure 28). Occasional sheetfloods from the craton deposited beds of the coarse sand sediment-type on the mudflats.

During the deposition of the upper Revett Formation, alluvial aprons and sandflats advanced across the study area from the southwest in four pulses, and deposited the intervals of flat-laminated sand in the upper Revett, and flat-laminated sand and even couples in the correlative interval of the Grinnell Formation (figure 28). A short-lived perennial lake on the upper Revett distal sandflats is recorded by a set of alpha couplets in the lower stratigraphic keybed at North Crow Creek (figure 28). Sheetfloods from the east crossed the mudflats and sandflats, depositing beds of the coarse sand sediment-type. With the final retreat of the upper Revett alluvial apron, the mudflats expanded to the southwest, and even and lenticular couplets accumulated to form the St. Regis Formation and much of the correlative interval of the Grinnell Formation (figure 28). The two stratigraphic intervals with interbeds of even couples (figure 28) reflect very minor alluvial pulses from the southwest, and the stratigraphic keybed which extends from North Crow Creek to Lion Creek records a large perennial lake in which a set of alpha couplets was deposited. Large alluvial aprons advanced into the Belt basin from the east, forming thick wedges of coarse sand that comprise nearly half of the thickness of the St. Regis correlative interval of the Grinnell Formation at Deep Creek. Sheetfloods commonly flowed across the exposed mudflats, ripping up mud polygons, and depositing beds of the coarse sand sediment-type as interbeds with the silt and clay of even and lenticular couplets derived from the southwest.

During the deposition of the Empire Formation, lenticular and even couplets were deposited on the mudflats. The Belt lake was larger and began to be perennial, as mud polygons from the mudflats were locally reworked into established beaches and shoals, forming beds of the sub-type 3 mudchip conglomerate sediment-type. Sheetfloods crossing the desiccated mudflats ripped up abundant mud polygons, and deposited them with silt and clay as beds of the sub-type 2 mudchip conglomerate sediment-type (Appendix E). The upward-increasing carbonate content of the Empire

Formation and the deposition of the overlying Middle Belt Carbonate records the beginning of the major transgression of the Belt lake across the basin (Winston, 1986c).

DISCUSSION

The preceding sedimentologic interpretations and lithofacies correlations clarify three points of controversy in the Ravalli Group: 1) the tectonic environment, 2) the significance of hematite/chlorite content, and 3) the tidal flat interpretation.

TECTONIC ENVIRONMENT

That vast quantities of fine sand, silt, and clay were transported into the Belt basin from the south and southwest during the deposition of the Ravalli Group, while only modest amounts of coarse sand were derived from the craton supports the inference that a landmass of continental proportions bounded the western margin of the Belt Basin (Cressman, 1985, 1988; Winston et. al, 1984; Winston, 1986b,c; Frost and Winston, 1987). The minor changes in the thickness of the Burke, Revett, and St. Regis Formations and their correlative parts of the Grinnell Formation (figure 29) across the area indicate that the eastern side of this intracratonic basin was tectonically stable, in contrast to the western and southern sides, in which syndepositional faults have been identified (Winston, 1986a). In addition, the minor changes in thickness and the remarkably detailed lithofacies correlations across the area (Figure 28; Appendix F) suggest that lateral displacement on Cenozoic normal faults between the measured sections (figure 4) is minor.

SIGNIFICANCE OF HEMATITE/CHLORITE CONTENT

Although hematite/chlorite content is a product of diagenesis, two aspects of the lithofacies correlations suggest that it often reflects some primary difference in the original chemistry of the sediments. One aspect is that coupleted intervals distinguished soley on the basis of their hematite/chlorite content correlate between measured sections. Another aspect is that defining the St. Regis-Spokane/Empire boundary at the change in the dominant color of couplets from red to green produces a remarkably consistent thickness of the St. Regis and the correlative interval of the Grinnell Formation (figure 28).

In a detailed sedimentologic study of the Spokane/Empire transition zone of west-central Montana, Whipple (1980) found that subaerial sedimentary structures such as sun-cracks characterized the red beds, whereas green beds reflect an absence of desiccation. In the Ravalli Group rocks of this study, lenticular couplets, which record long submergence, are more commonly chloritic (reduced), while even couplets, which are frequently sun-cracked, are more commonly hematitic (oxidized) Thus, the oxidation state of the sediments and enclosing pore waters is directly reflected in diagensis. Accordingly, green stratigraphic intervals within the Ravalli Group (figure 28) may represent topographically lower parts of the mudflats on which playas were often filled for relatively long periods of time, and the change to the dominant green color of the Empire Formation reflects a basinwide transgression of the Belt lake. Conversely, hematitic stratigraphic intervals within the Empire (figure 28) may record parts of the mudflats that were exposed. Although hematite/chlorite content, and thus color, are diagenetic features, the desiccated sediments of the mudflats probably did not generate great quantities of diagenetic fluids, and the clay layers probably restricted movement of fluids that were produced, thus preserving the original chemistry of the sediments on a regional scale. Although color can by no means be used as an absolute indicator of the dominant environmental conditions of coupleted intervals, neither should it be automatically discounted as insignificant, especially on a stratigraphic scale.

TIDAL FLAT INTERPRETATION

Lithofacies correlations and the sedimentologic interpretations indicate that the fine sediments of the

Grinnell are the distal facies of the entire Burke-Revett-St. Regis sedimentary prism. Thus, the Burke, Revett, and St. Regis Formations were not deposited simultaneously as different sub-environments of a large tide-dominated delta that prograded into the Belt basin, as suggested by Boyce (1973) and Mauk (1983).

Tidal mudflats are characterized by channels that funnel the daily flow both onto and off of the flats (Reineck and Singh, 1980). As tidal channels migrate laterally they erode the mudflats and deposit lateral accretion cross-beds (Reineck and Singh, 1980; Weimer et. al, 1982). Thompson's (1968, 1975) report that the tide flats bordering the estuary of the Colorado River lacked tidal channels is often cited to excuse the absence of tidal channel deposits in the Belt (Boyce, 1973; Mauk, 1983), but "in a subsequent study of the Colorado River delta, Meckel (1975) points out that the channels in the delta are in fact maintained by tidal flow, not by the emasculated Colorado River" (Winston, 1986c, p. 116). Thus, since the Ravalli Group mudflat deposits are totally devoid of tidal channel deposits, they cannot be reconciled with a tidal interpretation. Clearly, inundating <u>and</u> draining tidal mudflats as vast as those recorded in the Ravalli Group, diurnally, for millenia with no channellized flow seems impossible. In contrast, the sheetflooded mudflats of a playa-lacustrine basin are broad flat surfaces without channels (Hardie et. al, 1978), and numerous ancient playa-lacustrine mudflat deposits show no evidence of channelling (Smoot, 1983; Demicco and Kordesch, 1986; Flint, 1985; Cheadle, 1986; Hubert and Hyde, 1982).

Locally opposed cross-bed sets in the Ravalli Group, such as those in the cross-bedded sand and coarse sand sediment-types at North Crow Creek, have also been cited as clear evidence of tidal flow (Mauk, 1983), but recent studies by Mader and Teyssen (1985) and Alam and others (1985) conclude that bi-polar cross-beds are also characteristic of very low gradient fluvial systems. The sets of the cross-bedded sand sediment-type interstratified with laterally extensive beds of the flat-laminated sand sediment-type, and of sets of the cross-bedded coarse sand sub-type with unchannelled mudflat deposits of even and lenticular couplets are inconsistent with a tidal interpretation, but strongly support the playa model proposed here and by Winston (1986c).

Synthesis of the sediment-type facies tracts (figure 32) reveals that each ephemeral, terrestrial sheetflood deposited an individual basinward-thinning and -fining sedimentary layer (figure 31). Clearly, Unit B, and the Burke, Revett, and St. Regis formations and their respective correlative intervals of the Grinnell

Formation record an intracratonic alluvial apron-sandflat-playa mudflat system.

Chapter 5

CONCLUSIONS

Lithofacies correlations from the Mission Range, through the Swan Range, and into the Flathead Range indicate that the entire Burke-Revett-St. Regis sequence correlates with the Grinnell (Spokane) Formation (figure 30), not just the St. Regis Formation as proposed by Harrison (1972) (figure 5).

A major, informal stratigraphic unit dominated by fine- to very fine-grained quartzite occurs between the Prichard Formation, and the Burke Formation and the correlative interval of the Grinnell (Spokane) Formation and is called Unit B (figure 30).

The unit of dominantly green argillite with a few beds of medium- to coarse-grained quartz arenite that occurs between Unit B and the Prichard Formation in the Flathead Range is designated informal Unit A and is tentatively correlated eastward with the uppermost member of the Appekunny Formation of Glacier National Park, member 5 (Whipple et. al, 1984). Unit A pinches out westward, so that in the Swan Range Unit B directly overlies the Prichard Formation (figure 29). Thus, the Appekunny Formation does not correlate with the Burke Formation, as proposed by Harrison (1972) (figure 5) and is not part of the Ravalli Group. Instead, Unit A correlates with the Prichard Formation or with Unit B.

The Ravalli Group and Unit B are interpreted to record intracratonic, alluvial apron-sandflat-playa mudflat environments. Episodic floods from a western continent flowing down braided streams across vast alluvial aprons that sloped into the basin from the south and southwest (figure 32) broadened and shallowed progressively downstream to sheetfloods as the slope of the apron decreased. Isolated bars of fine- to very fine-grained sand migrating in the upper regime, braided flow within proximal sheetfloods deposited sets of the cross-bedded sand sediment-type interstratified with beds of the flat-laminated sand sediment-type. The sheetfloods thinned and spread out across the toe of the alluvial apron and onto sandflats, depositing beds of the flat-laminated sand sediment-type, and the lower layers of the even couple sediment-type. Flowing out across the mudflats, the sheetfloods continued to thin and decelerate, depositing the lower layers of even couplets. As flood waters ponded and playa lakes spread across the mudflats, clay settled from suspension, forming clay caps. Some playa bottoms dried, and the clay layers were soon desiccated and cracked. Areas of the playa that remained submerged were locally reworked by waves into the lenticualr couplet sediment-type. Sheetfloods that entered rare perennial lakes on the distal sandflats formed turbid plumes that settled out to form alpha couplets. Occasional sheetfloods from the east flowed across localized alluvial aprons of medium- to coarse-grained sand and spread out across the exposed mudflats, depositing beds of the coarse sand sediment-type. Rare eastern sheetfloods crossed the mudflats and reached the sandflats and even the toe of the fine-grained alluvial apron on the western side of the basin, so that beds of the coarse sand sediment-type were intercalated with beds of the even couple or flat-laminated sand sediment-types. Thus, the sand wedges of Unit B, the lower Revett and its correlative Grinnell interval, and the upper Revett and its correlative Grinnell interval record alluvial advances into the Belt basin, while the Burke, middle Revett, and St. Regis, and their correlative parts of the Grinnell Formation record episodes of alluvial retreat and the expansion of the playa mudflats to the south and southwest.

REFERENCES CITED

Alam, M.M., Crook, K.A.W., and Taylor, G., 1985, Fluvial herring-bone cross-stratification in a modern tributary mouth bar, Coonamble, New South Wales, Australia: Sed., v. 32, p. 235-244.

Alleman, D.G., 1983, Stratigrahy and sedimentation of the Precambrian Revett Formation, northwest Montana and northern Idaho: unpub. M.S. thesis, University of Montana, Missoula, 103 p.

Allen, J.R.L, 1973, Phase differences between bed configurations and flow in natural environments, and their geological significance: Sed., v. 20, p. 323-329.

_____, 1984a, Sedimentary structures, their character and physical basis: v. 1 and 2, Elsevier, New York.

_____, 1984b, Parallel lamination developed from upper-stage plane beds: a model based on the larger coherent structures of the turbulent boundary layer: Sed. Geol., v. 39, p. 227-242.

Allen, J.R.L. and Collinson, J.D., 1974, Reaction, relaxation and lag in natural sedimentary systems: general principles, examples and lessons: Earth Sci. Rev., v. 10, p. 263-342.

Allen, J.R.L. and Leeder, M.R., 1980, Criteria for the instability of upper stage plane beds: Sedimentology, v. 27, p. 209-217.

Ashley, G.M., Southard, J.B., and Boothroyd, J.C., 1982, Deposition of climbing ripple beds: a flume simulation: Sedimentology, v. 29, p.67-79.

Banerjee, I., 1977, Experimental study on the effect of deceleration on the vertical sequence of sedimentary structures in silty sediments: J. Sed. Pet., v. 47, p. 771-783.

Blair, T.C., 1987, Sedimentary processes, vertical stratification sequences, and geomorphology of the Roaring River alluvial fan, Rocky Mountain National Park, Colorado: J. Sed. Petr., v. 57, n. 1, p. 1-18.

Blodgett, R.H. and Stanley, K.O., 1980, Stratification, bedforms, and discharge relations of the Platte braided river system, Nebraska: J. Sed. Pet., v. 50, no. 1, p. 139-148.

Bowden, T.D., 1977, Depositional processes and environments within the Revett Formation, Precambrian Belt basin, northwestern Montana and northern Idaho: Unpub. M.S. thesis, University of California, Riverside, 161 p.

Boyce, R.L., 1973, Depositional systems in the Ravalli Group: a conceptual model and possible modern analogue: Belt Symposium v.1, Idaho Bur. Mines and Geology, p. 139-158.

Burst, J.F., 1965, Subaqueouly formed shrinkage cracks in clay: J. Sed. Pet., v. 35, p. 348-353.

Cant, D.J. 1978a, Fluvial processes and facies sequences in the sandy braided South Sascatchewan River, Canada: Sedimentology, v. 25, p. 625-648.

______, 1978b, Bedforms and bar types in the south Saskatchewan River: Jour. Sed. Petrology, v. 48, p. 1321-1330.

Cant, D.J. and Walker, R.G., 1976, Development of a braided fluvial facies model of the Devonian Battery Point Sandstone, Quebec: Can. J. Earth Sci. v. 13, p. 102-119.

Cheadle, B.A., 1986, Alluvial-playa sedimentation in the lower Keweenawan Sibley Group, Thunder Bay District, Ontario: Can. J. Earth Sci., v. 23, p. 527-542.

Cheel, R.J., and Middleton, G.V., 1986, Horizontal laminae formed under upper flow regime plane bed conditions: J. Geol., v. 94, p. 489-504.

Collins, J.A. and Smith, L., 1977, Genesis of cupriferous quartz arenite cycles in the Grinnell Formation (Spokane Equivalent), Middle Proterozoic Belt-Purcell Supergroup, Eastern Rocky Mountains, Canada: Bull. Can. Petrol. Geol., v. 25, no. 4, p. 713-735.

Connor, J.J., Reynolds, M.W., and Whipple, J.W., 1984, Stratigraphy of the Ravalli Group, Belt basin, Montana and Idaho (abstr.): Belt Symposium II proceedings, 1983, Montana Bur. Mines and Geol., Special Publication 90, p. 13-15.

Constenious, K.N., 1981, Stratigraphy, sedimentation, and tectonic history of the Kishenehn basin, northwestern Montana: Master's thesis, University of Wyoming, Laramie, 116 p.

Costello, W.R. and Southard, J.B., 1981, Flume experiments on lower-flow regime bed forms in coarse sand: J. Sed. Pet., v. 51, no. 3, p. 849-864.

Cressman, E.R., 1985, The Prichard Formation of the Lower Part of the Belt Supergroup (Middle Proterozoic), near Plains, Sanders County Montana: U.S.G.S. Bull. 1553, 64 p.

Cronin, C., Kuhn, J.A., and Winston, Don, 1986, Stratigraphy of the Ravalli Group (Middle Proterozoic Belt Supergroup) between the Coeur D'Alene District, Idaho, and Glacier National Park, Montana: G.S.A. Abstr. with Progr., v. 18, n. 5, p. 349.

Crowley, K.D., 1983, Large-scale bed configurations (macroforms), Platte River basin, Colorado and Nebraska: primary structures and formative processes: G.S.A. Bull, v. 94, p. 117-133.

Cummins, W.A., 1956, Some sedimentary structures from the lower Keuper Sandstones, Liverpool and Manchester Geol. Journal, v. 2, p. 37-43.

Curray, J.R., 1956, The analysis of two dimensional orientation data: Jour. Geology, v. 64, p. 117-131.

Davis, W.M., 1938, Sheetfloods and streamfloods: G.S.A. Bull., v.49, p. 1337-1416.

Demicco. R.V. and Kordesch, E.G., 1986, Facies sequences of semi-arid closed basin: the Lower Jurassic East Berlin Formation of the Hartford Basin, New England, U.S.A.: Sedimentlogy, v. 33, p. 107-118.

Donovan, R.N., and Foster, R.J., 1972, Subaqueous shrinkage cracks from the Caithness Flagstone Series (Middle Devonian) of northeast Scotland: J Sed. Pet., v. 42, p. 309-317.

Earhart, R.L., Mudge, M.R., and Connor, J.J., 1984, Belt Supergroup lithofacies in the Northern Disturbed Belt, northwest Montana: Montana Geol. Soc. Guidebook, 1984 Field Conf., Northwest Montana and Adjacent Canada, p. 51-56.

Eugster, H. P. and Hardie, L. A., 1975, Sedimentation in an ancient playa-lake complex: The Wilkins Peak Member of the Green River Formation of Wyoming: Geol. Soc. Am. Bull., v. 86, p. 319-334.

Fahnestock, R.K. and Haushild, W.L., 1962, Flume studies of the transport of pebbles and cobbles on a sand bed: Bull. Geol. Soc. Am., v. 78, p. 1431-1436.

Flint, S., 1985, Alluvial fan and playa sedimentation in an arid closed basin: the Pacencia Group, Antofagasta Province, Chile: J. Geol. Soc. Lond., v. 142, p. 533-546.

_____, 1987, Diagenesis of Tertiray playa sandstones of Northern Chile; implications for Andean uplift and metallogeny: Sedimentology, v. 34, p. 11-29.

Freeman, W. and Winston, Don, 1987, A quartz arenite blanket at the base of or below the Middle Proterozoic Belt Supergroup?, Montana and Idaho: G.S.A. Abstr. with Progr., v. 19, no. 5, p. 276.

Frost, C.D. and Winston, D., 1987, Nd isotope systematics of coarse- and fine-grained sediments: examples from the Belt-Purcell Supergroup: J. of Geology, v. 95, p. 309-327.

Gibson, Russel, 1948, Geology and ore deposits of the Libby quadrangle, Montana: U.S. Geol. Survey Bull. 956, 131p.

Gosh, J.K., Mazumder, B.S., Saha, M.R., and Sengupta, S., 1986, Deposition of sand by suspension currents: experimental and theoretical studies: J. Sed. Pet., v. 56, p. 57-66.

Greene, S.E., 1984, Stratigraphy and sedimentation of the Revett Formation, Precambrian Belt Supergroup, Shoshone and Bonner Counties, northern Idaho: unpub. M.S. thesis, University of Idaho, Moscow, 177p.

Grotzinger, J.P., 1986, Shallowing-upward cycles of the Wallace Formation, Belt Supergroup, northwestern Montana and northern Idaho: in Roberts, S.M., ed., Belt Supergroup: a guide to Proterozoic rocks of western Montana and adjacent areas, Montana Bur. of Mines and Geol. Spec. Pub. 94, p. 143-160.

Hamblin, W.K. 1962a, X-ray radiography in the study of structures in homogeneous sediments: J. Sed. Pet. v. 32, p. 201.

_____, 1962b, Staining and etching techniques for studying obscure structures in clastic rocks: J. Sed. Pet., v. 32, p. 350.

Hardie, L.A., Smoot, J.P., and Eugster, H.P., 1978, Saline lakes and their deposits: a sedimentologic approach, in Matter, M.A., and Tucker, M.E., eds., Modern and Ancient Lake Deposits, Spec. Publs. Int. Ass. Sediment., v. 2: Oxford, Blackwell Scientific Publications, p. 7-41.

Harms, J.C. and Fahenstock, R.K., 1965, Stratification, bed forms, and flow phenomena (with an example from the Rio Grande): SEPM Spec. Pub. 12, p. 84-115.

Harms, J.C., Southard, J.B., Spearing, D.R., and Walker, R.G., 1975, Depositional Environments as Interpreted from Primary Sedimentary Structures and stratification Sequences.": SEPM Short Course Lecture Notes 2, 161 p.

Harms, J.C., Southard, J.B., and Walker, R.G., 1982, Structures and sequences in clastic rocks: SEPM Short Course No. 9, 249 p.

Harrison, J.E., and Jobin, A.B., 1963, Geology of the Clark Fork quadrangle, Idaho-Montana: U.S. Geol. Survey Bull., 1141-k, 38 p.

Harrison, J.E., 1972, Precambrian Belt Basin of northwestern United States --- its geometry, sedimentation, and copper occurrences: Geol. Soc. of America Bull., v.83, no.5, p. 1215-1240.

Harrison, J.E., Griffs, A.B., and Wells, J.D., 1974, Tectonic features of the Precambrian Belt basin and their influence on post-Belt structures: U.S. Geol. Survey Prof. Paper 866, 15 p.

Harrison, J.E., Kleinkopf, M.D., and Wells, J.D., 1980, Phanerozoic thrusting in Proterozoic Belt rocks, northwestern United States: Geology, v. 8, p. 407-411.

Harrison, J.E., Cressman, E.R., and Whipple, J.W., 1983, Preliminary geologic and structural map of part of the Kalispell 1 x 2 quadrangle, Montana: U.S. Geol. Survey, open File report 83-502.

Harrison, J.E., Griggs, A.B., and Wells, J.D., 1986, Geologic and structure maps of the Wallace 1 x 2 quadrangle, Montana and Idaho: Montana Bureau of Mines and Geolocy Montana Atlas Series 4-a, Scale 1:250,000.

Herndon, S.D., 1983, Diagenesis and metamorpnhism in the Revett quartzite, Middle Proterozoic Belt: unpub. M.S. thesis, Univ. of Montana, Missoula, 66 p.

High, L.R., Jr., and Picard, M.D., 1974, Reliability of cross-stratifcation types as paleocurrent indicators in fluvial rocks: Jour. Sed. Petrol., v. 44, p. 158-168.

Hobbs, S.W., Griggs, A.B., Wallace, R.E., and Campbell, A.B., 1965, Geology of the Coeur d'Alene district, Shoshone County, Idaho: U.S. Geol. Survey Prof. Paper 478, 139 p.

Hogg, S.E., Sheetfloods, sheetwash, sheetflow, or...?: Earth. Sci. Rev., v. 18, p. 59-76.

Horodyski, R.J., 1983, Sedimentary geology and stromatolites of the Middle Proterozoic Belt Supergroup, Glacier National Park, Montana: Precambrian Research, v.20, p.391-425.

Hrabar, S.V., 1973, Deep-water sedimentation in the Ravalli Group (Late Precambrian Belt Megagroup), northwestern Montana: in Belt Symposium 1973, Volume 2, University of Idaho, p. 67-81.

Hubert, J.F., and Hyde, M.G., 1982, Sheet-flow deposits of graded beds and mudstones on an alluvial sandflat-playa system: Upper Triassic Blomidon redbeds, St. Mary's Bay, Nova Scotia: Sedimentology, v. 29, n. 4, p. 457-474.

Hunter, R.E., 1985, Subaqueous sand-flow cross-strata: J. Sed. Pet., v. 55, no. 6, p. 886-894.

Hunter, R.E. and Kocureck, G., 1986, An experimental study of subaqueous slipface deposition: J. Sed. Pet., v. 56, no. 3, p. 387-394.

Johns, W.M., 1970, Geology and mineral deposits of Lincoln and Flathead Counties, Montana: Montana Bureau of Mines and Geology Bulletin 42, 66p.

Jones, C.M., 1977, Effects of varying discharge regimes on bed-form sedimentary structures in modern rivers: Geology, v. 5, p. 567-570.

Jopling, A.V., 1965, Hydraulic factors controlling the shape of laminae in laboratory deltas: Jour. Sed. Pet., v. 35, n. 4, p. 777-791.

Jopling, A.V., and Walker, R.G., 1968, Morphology and origin of rippledrift cross-lamination, with examples from the Pleistocene of Massachusetts: Jour. Sed. Petrol., v. 38, P. 971-984.

Kuenen, PH.H., 1966a, Experimental turbidite lamination in a circular flume: J. Geol. v. 74, p. 523-545.

_____, 1966b, Matrix of turbidites: experimental approach: Sedimentology, v. 7, p. 267-297.

_____, 1967, Emplacement of flysch-type sand beds: Sedimentology, v. 9, p. 203-243.

Kuenen PH.H. and Sengupta, S., 1970, Experimental marine suspension currents, competency and capacity: Geologie en Mijnbouw, v. 49, p. 89-118.

Kuhn, J.A., 1986, The stratigraphy and sedimentology of the Middle Proterozoic Grinnell Formation, Glacier National Park and the Whitefish Range, N.W. Montana: unpub. M.S. thesis, Univ. of MT, Missoula, MT, 113p.

Lemoine, S.R. and Winston, Don, 1986, Correlation of the Snowslip and Shepard formations of the Cabinet Mountains with upper Wallace rocks of the Coeur D'Alene Mountains, western Montana: in Roberts, S.M., ed., Belt supergroup: a guide to Proterozoic rocks of western Montana and adjacent areas, Montana Bur. Mines and Geol., Spec. Pub. 94, p. 161-168.

Lindholm, R.C., 1982, Flat Stratification: Two ancient examples: J. Sed. Pet., V. 52, p. 227-231.

Loos, K.D., 1985, A sedimentological study of the transition between the Prichard Formation and Ravalli Group, Salish Mountains, northwest Montana: unpub. M.S. thesis, Univ. of Cincinnati, 153p.

Lowe, D.R., 1975, Water escape structures in coarse-grained sediments: Sedimentology, v. 22, p. 157-204.

_____, 1982, Sediment gravity flows II: depositional models with special reference to the deposits of high-density turbidity currents: J. Sed. Pet., v. 52, p. 279-297.

_____, 1986, Suspended sediment fallout rate as an independent variable in bedform and structure analysis (abstr.): Abstracts from the 12th International Sedimentologic Congress, August, 1986, Canberra, Australia.

Mader, D. and Teyssen, T., 1985, Palaeoenvironmental interpretation of fluvial red beds by statistical analysis of palaeocurrent data: examples from the Buntsandstein (Lower Triassic) of the Eifel and Bavaria in the German Basin (Middle Europe): Sed. Geol., v. 41, p. 1-74.

Mauk, J.L., 1983, Stratigraphy and sedimentation of the Proterozoic Burke and Revett Formations, Flathead Reservation, western Montana: M.S. thesis, University of Montana, Missoula.

McBride, E.F., Shepard, R.G., and Crawly, R.A., 1975, Origin of parallel near-horizontal laminae by migration of bed forms in a small flume: Jour. Sed. Pet., v. 45, n.1, p. 132-139.

McGee, E.J., 1897, Sheetflood erosion: Geol. Soc. Am. Bull., v. 8, p. 87-112.

McKee, E.D., 1957, Flume experiments on the production of stratification and cross-stratification: J. Sed. Pet., v. 27, no. 2, p. 129-134.

McKee, E.D., Crosby, E.J., and Berryhill, H.L., 1967, Flood deposits, Bijou Creek, Colorado, June, 1956: Jour. Sed. Petrol., v. 37, p. 839-851.

McMechan, M.E., 1981, The Middle Proterozoic Purcell supergroup in the southwestern Rocky and southeastern Purcell Mountains, British Columbia and the initiation of the Cordilleran Miogeocline, southern Canada and adjacent United States: Bull. Canad. Petrol. Geol., v. 29, no. 4, p. 583-621.

Meckel, L.D., 1975, Holocene sand bodies in the Colorado delta area, northern Gulf of California, in Deltas: Models for Exploration, M.L. Broussard, (ed.): Houston Geological Society, Houston, Texas, p. 239-265.

Miall, A.D., 1974, Paleocurrent analysis of alluvial sediments; a discussion of directional variance and vector magnitude: Jour. Sed. Petrol., v. 44, p. 1174-1185.

_____, 1976, Paleocurrent and paleohydrologic analysis of some vertical profiles through a Cretaceous braided stream deposits, Banks Island, Arctic Canada: Sedimentology, v. 23, p. 459-483.

_____, 1977, A review of the braided-river depositional environment: Earth Sci. Rev., v. 13, p. 1-62.

_____, 1985, Architectural-element analysis: a new method of facies analysis applied to fluvial deposits: Earth-Science Reviews, v. 22, p. 475-504.

Middleton, G.V., 1967, Experiments on density and turbidity currents III: deposition of sediments: Can. J. Earth Sci., v. 4, p. 475-504.

Mills, P.C., 1983, Genesis and diagnostic value of soft-sediment deformation structures -- a review: Sed. Geol., v. 35, p. 83-104.

Molenaar, N., 1986, The interrelation between clay infiltration, quartz cementation, and compaction in Lower Givetian terrestrial sandstones, Northern Ardennes, Belgium: J. Sed. Pet., v. 56, no. 3, p. 359-369.

Moss, A.J., 1972, Bed-load sediments: Sedimentology, v. 18, p. 159-219.

Moss, A.J., Walker, P.H., and Hutka, J., 1980, Movement of loose, sandy detritus by shallow water flows: an experimental study: Sed. Geol., v. 25, p. 43-66.

Mudge, M.R., 1970, Origin of the disturbed belt in northwestern Montana, Geol. Soc. America Bull., v. 81, p.377-392.

_____, 1977, General geolgoy of Glacier National Park and adjacent areas, Montana: Bull. Canad. Petrol. Geol., v. 25, no. 4, p. 736-751.

Mudge, M.R., Earhart, R.L., Whipple, J.W., and Harrison, J.E., 1983, Geologic and structure maps of the Choteau 1 by 2 Quadrangle, northwestern Montana: Montana Bureau of Mines and Geology Montana Atlas Series 3-A, Scale 1:250,000.

Muir, M., Lock, D., and Von Der Borch, C., 1980, The Coorong Model for penecontemporaneous dolomite formation in the Middle Proterozoic McArthur Group, Northen Territory, Australia: in, Zenger, D.H., Dunham, J.B., and Ethington, R.L., eds., Concepts and Models of Dolomitization, SEPM Spec. Pub. 28.

Obradovich, J.D., Zartman, R.E., and Peterman, Z.E., 1984, Update of the Geochronology of the Belt Supergroup: in Hobbs, S.W., The Belt, Montana Bur. Mines and Geol. Spec. Pub. 90, p. 82-84.

Picard, M.D., and High, L.R., Jr., 1973, Sedimentary structures of ephemeral streams: Elsevier, New York, 233 p.

Plummer, P.S. and Gostin, V.A., 1981, Shrinkage Cracks: desiccation or synaeresis?: J. of Sed. Petr., v. 51, n. 4, p. 1147-1156.

Potter, P.E. and Pettijohn, F.L., 1977, Paleocurrents and basin analysis: Springer-Verlag, New York, 306p.

Price, R.A., 1964, The Precambrian Purcell System in the Rocky Mountains of southern Alberta and British Columbia: Bull. of Canad. Petrol. Geol., v. 12, Special Issue (Guidebook), p. 399-426.

Price, R.A. and Kluvver, H.M., 1974, Structure of the Rocky Mountains of the Glacier-Waterton Lakes National Park area: in Rock Mechanics: the American Northwest, International Society for Rock Mechanics, 3rd Congress Expedition Guide, Special Publication, Experiment Station, college of Earth and Mineral Sciences, The Pennsylvania State University, University Park, Pa., 4p.

Rahn, P.H., 1967, Sheetfloods, streamfloods, and the formation of sediments: Ann. Am. Assoc. Geogr., v. 57, p. 593-604.

Ransome, F.L., 1905, Ore deposits of the Coeur d'Alene district, Idaho: U.S. Geol. Survey Bull. 260., p. 274-303.

Ransome, F.L. and Calkins, F.C., 1908, The geology and ore deposits of the Coeur d'Alene district, Idaho: U.S. Geol. Survey Prof. Paper 62, 203 p.

Raup, O.B., Earhart, R.L., Whipple, J.W., and Carrara, P.E., 1983, Geology along Going-to-the-Sun Road, Glacier National Park, Montana, a self guided tour for motorists: Glacier Natural History Association, West Glacier, Montana, 62 p.

Reineck, H.E. and Singh, I.B., 1971, Genesis of laminated sand and graded rhythmites in storm-sand layers of shelf mud: Sedimentology, v. 18, p. 123-128.

Reineck, H.E. and Singh, I.B., 1980, Depositional Sedimentary Environments 2nd ed., New York, Springer-Verlag, 549 p.

Reynolds, M.W., 1984, Tectonic setting and development of the Belt Basin, northwestern United States (Abstr.): in Hobbs, S.W., ed., The Belt, Montana Bur. Mines and Geol. Spec. Pub. 90, p. 44-46.

Ross, C.P., 1959, Geology of Glacier National Park and the Flathead region, northwestern Montana: U.S. Geol. Survey Prof. Paper 296, 125 p.

Rubin, D.M. and Hunter, R.E., 1982, Bedform climbing in theory and nature: Sedimentology, v. 29, v. 121-138.

Sears, J.W., 1986, A two thrust-slab model for the central Montana-Idaho overthrust Belt, and its relationship to the Rocky Mountain foreland: (abstr.) Geol. Soc. Am. Abstracts with Programs, v. 18, no. 5, p. 412.

Simons, D.P., Richardson, E.V., and Haushild, W.L., 1963, Some effects of fine sediment on flow phenomena: U.S.G.S. Water Supply Paper 1498-G, 47 p.

Simons, D.P., Richardson, E.V., and Nordin, C.F., 1965, Sedimentary structures generated by flow in alluvial channels: SEPM Spec. Pub. 12, p. 34-52.

Singh, I.B., 1972, On the bedding in the natural-levee and the point-bar deposits of the Gomti River, Uttar Pradesh, India: Sed. Geol., v. 7, p. 309-317.

Smith, A.G. and Barnes, W.C., 1966, Correlation of and facies changes in the carbonaceous, calcareous, and dolomitic formation of the Precambrian Belt-Purcell Supergroup: Geol. Soc. Am. Bull., v. 77, no. 12, p. 1399-1426.

Smith, N.D., 1970, The braided stream depositional environment: comparison of the Platte River with some Silurian Clastic Rocks, North-Central Appalachians: Geol. Soc. of America Bull., v. 81, p. 2993-3014.

_____, 1971, Pseudo-planar stratification produced by very low amplitude sand waves: J. Sed. Pet., v. 41, p. 69-73.

_____, 1972, Some sedimentologic aspects of planar cross-stratification in a sandy braided river: Jour. Sed. Petrol., v. 42, p. 624-634.

Smoot, J.P., 1983, Depositional subenvironments in an arid closed basin; the Wilkins Peak Member of the Green River Formation (Eocene), Wyoming, U.S.A.: Sedimentology, v. 30, p. 801-827.

_____, 1985, Subaerial exposure criteria as seen in modern playa mudcracks: Unpub. manuscript, 11 pgs.

Smoot, J.P. and Katz, S.B., 1985, A comparison of modern playa mudflat fabrics to cycles in the Triassic Lockatong Formation of New Jersey and Pennsylvania: Unpub. manuscript.

Sneh, A., 1983, Desert stream sequences in the Sinai Peninsula: J. Sed. Pet., v. 53, no. 4, p. 1271-1279.

Stear, W.M., 1985, Comparison of the bedform distribution and dynamics of modern and ancient sandy ephemeral flood deposits in the southwestern Karoo region, South Africa: Sed. Geol., v. 45, p. 209-230.

Surdam, R.C. and Wofbauer, C.A., 1975, Green River Formation, Wyoming: A playa-lake complex: Geol. Soc. Am. Bull., v. 86, p. 335-345.

Thompson, R.W., 1968, Tidal flat sedimentation on the Colorado River delta: G.S.A. Memoir No. 107, 133 p.

Thompson, R.W., 1975, Tidal-flat sediments of the Colorado River delta, northwestern Gulf of Califonia, in Ginsburg, R.N., ed., Tidal Deposits, New YOrk, Springer-Verlag, p. 57-66.

Tunbridge, I.P., 1981, Sandy high-energy flood sedimentation - some criteria for recognition, with an example from the Devonian of S.W. England: Sed. Geol., v. 28, p. 79-95.

_____, 1983, Alluvial fan sedimentation of the Horseshoe Park flood, Colorado, U.S.A., July 15th, 1982: Sed. Geol., v. 36, p. 15-23.

_____, 1984, Facies model for a sandy ephemeral stream and clay playa complex; the Middle Devonian Trentishoe Formation of North Devon, U.K.: Sedimentology, v. 31, p. 697-715.

Turner, P., 1980, Continental Red beds: Elsevier, New York, 562 p.

Visher, G.E. and Cunningham, R.D., 1981, Convolute laminations -- A theoretical Analysis: Example of a Pennsylvanian sandstone: Sed. Geol., v. 28, p. 175-188.

Walcott, C.D., 1899, Pre-Camgrian fossiliferous formations: Geol. Soc. of America Bull., v. 10, p. 199-244.

_____, 1906, Algonkian formations of northwestern Montana: Geol. Soc. of America Bull., v. 17, p. 1-28.

Weimer, R.J., Howard, J.D., and Lindsay, D.R., 1982, Tidal flats and associated tidal channels: in Sandstone Depositional Environments, American Assoc. Petrol. Geologists, Memoir 31, Scholle, P.A. and Spearing, D., (eds.).

Wheeler, H.E. and Mallory, V.S., 1953, Designation of stratigraphic units: American Assoc. of Petrol. Geologists Bull., v. 37, n. 10, p. 2407-2421.

Walker, T.R., Waugh, B., and Crone, A.J., 1978, Diagenesis in first cycle desert alluvium of Cenozoic age, southwestern United States and northwestern Mexico: Bull. Geol. Soc. Am., v. 89, p. 19-32.

Whipple, J.W., 1980, Depositional environment of the Middle Proterozoic Spokane Formation - Empire Formation Transition Zone, west-central Montana: U.S. Geol. Survey Open File Rept. 80-1232, 98 p.

Whipple, J.W., Connor, J.J., Raup, O.B., and McGimsey, R.G., 1984, Preliminary report on the stratigraphy of the Belt Supergroup, Glacier National Park and adjacent Whitefish Range, Montana: Montana Geol. Soc. Guidebook, 1984 Field Conf., Northwest Montana and adjacent Canada, p. 33-50.

White, B.G., and Winston, Don, 1982, The Revett-St. Regis "Transition Zone" near the Bunker Hill mine, Coeur d'Alene District, Idaho: Idaho Bur. Mines and Geol. Bull. 24, p.25-30.

White, B.G., Winston, Don, and Jacob, P. 1977, The Revett Formation near Kellogg, Idaho: Geol. Soc. America (abs.): v. 9, p. 733.

Williams, G.E., 1971, Flood deposits of the sand-bed ephemeral streams of central Australia: Sedimentology, v. 17, p. 1-40.

Willis, Bailey, 1902, Stratigraphy and structure, Lewis and Livingston ranges, Montana: Geol. Soc. of America Bull., v. 13, p. 305-352.

Wingerter, J.H., 1982, Depositional environment of the Revett Formation, Precambrian Belt Supergroup, northern Idaho and northwestern Montana: unpub. M.S. thesis, Eastern Washington University, 119 p.

Winston, Don; Woods, M., and Byer, G.B., 1984, The case for an intracratonic Belt-Purcell Basin: tectonic, stratigraphic and stable isotope considerations: Montana Geol. Soc. Guidebook, 1984 Field Conf., Northwest Montana and Adjacent Canada, p. 33-50.

Winston, Don, 1986a, Tectonics and sedimentation of the Middle Proterozoic Belt basin, and their influence on Cretaceous compression and Cenozoic extension in western Montana and Northern Idaho: <u>in</u> Peterson, J.A., ed., Paleotectionics and Sedimentation, A.A.P.G. Memoir 41, p. 87-118.

Winston, Don, 1986b, Stratigraphic correlation and nomenclature of the Middle Proterozoic Belt Supergroup, Montana, Idaho and Washington: in Roberts, S.M., ed., Belt Supergroup: a guide to Proterozic rocks of western Montana and adjacent areas, Montana Bur. Mines and Geol., Spec. Pub. 94, p. 69-84.

Winston, Don, 1986c, Sedimentology of the Ravalli Group, middle Belt carbonate and Missoula Group, Middle Proterozoic Belt Supergroup, Montana, Idaho and Washington: in Roberts, S.M., ed., Belt Supergroup: a guide to Proterozic rocks of western Montana and adjacent areas, Montana Bur. Mines and Geol., Spec. Pub. 94, p. 85-124.

Winston, Don, 1986d, Middle Proterozoic tectonics of the Belt basin, western Montana and northern Idaho: in Roberts, S.M., ed., Belt Supergroup: a guide to Proterozic rocks of western Montana and adjacent areas, Montana Bur. Mines and Geol., Spec. Pub. 94, p. 245-258.

Woods, Marvin, O., 1985, Depositional subenvironments in a closed basin: the Shepard Formation (Middle Proterozoic Belt Supergroup), southern Mission, Swan, and Lewis and Clark Ranges, Montana: M.S. thesis, University of Montana, Missoula, 215p.

Cronin, Kuhn, and Winston, 1986,
G.S.A. Abstr. with Progr., v. 18, n. 5, p. 349.

STRATIGRAPHY OF THE RAVALLI GROUP (MIDDLE PROTEROZOIC BELT SUPERGROUP) BETWEEN THE COEUR D'ALENE DISTRICT, IDAHO, AND GLACIER NATIONAL PARK, MONTANA

CRONIN, Chistopher, KUHN, Jeffrey A., and WINSTON, Don, (all authors) Geology Dept., Univ. of Montana, Missoula, MT 59812

The Burke Formation (lower Ravalli) of the Coeur d'Alene district is currently correlated with the Appekunny Formation of the Lewis Range in Glacier National Park; the St. Regis Formation (upper Ravalli) of the Coeur d'Alene district is correlated with most of the Grinnell Formation of Glacier National Park; while the Revett Formation (middle Ravalli) of the Coeur d'Alene district pinches out eastward within the Appekunny-Grinnell sequence. In the eastern part of the basin the names Greyson and Spokane have been used somewhat interchangeably with Appekunny and Grinnell respectively.

Measured sections from the Mission, Swan, Whitefish, Clark, and Lewis ranges indicate that the Burke Formation of the Coeur d'Alene district extends eastward as siltite, fine quartzite, and argillite intervals which may correlate partially with the uppermost Appekunny, but which correlate mostly with the lower Grinnell in the Lewis Range. The lower Revett quartzite passes eastward to purple siltite and argillite mixed with beds of coarse quartzite within the lower Grinnell of the Lewis Range. The middle Revett siltite of the Coeur d'Alene district passes to red argillite and mudstone of the middle part of the Grinnell in the Lewis Range, and the upper Revett quartzite units pass to very fine quartzite, siltite and argillite of the Lewis Range mixed with beds of coarse quartzite that increase upward. The St. Regis Formation passes eastward to argillite and coarse quartzite beds in the Swan Range, correlating with only the uppermost siltite, argillite and coarse quartzite of the Grinnell in the Lewis Range. Correlation of the Burke, Revett, and St. Regis with the uppermost Appekunny and the Grinnell may simplify formal stratigraphic usage.

APPENDIX B

PALEOCURRENT METHODOLOGY

Paleocurrent analysis of the cross-bedded sand sediment-type and the cross-bedded sub-type of the coarse sand sediment-type in the Revett Formation at North Crow Creek utilized measurements of tabular cross-bed sets. A 4"x6" plexiglass panel was laid directly on the steepest part of an exposed foreset plane and its dip and strike were measured with a Brunton compass. In most cases the attitude of cross-strata in a set remained esentially constant, so only one foreset was measured, but where attitudes varied, several foresets were measured at 1-3 meter spacings across the outcrop. The attitude of regular bedding was also recorded. Data was taken from all tabular cross-bed sets encountered in the measured section. In order to increase this data base, the Revett was scouted up another route, with measurements made wherever possible. Height in the section, corrected foreset attitude, set thickness, foreset shape, mode of occurrence, vector mean azimuth and consistency ratio for each set are listed in tables 1 and 2.

To correct the effects that tectonic tilting has had on the cross-strata's original attitude, the bedding containing each cross-bed set was returned to horizontal. For each cross-bed attitude measurement, a Schmidt equal-area stereonet was used for the following procedure

(Potter and Pettijohn, 1977): 1) the poles to bedding and cross-bedding are plotted, 2) with the bedding pole on the E-W line, the cross-bed pole is "moved" along its small circle in the direction and by the number of degrees required to return the bedding pole to the stereonet center, 3) the restored cross-bed pole is plotted and the corrected foreset strike and dip is recorded. However, because possible rotational aspects of Mesozoic thrust faulting and Cenozoic normal faulting can not be corrected-for given the imprecise knowledge of the structures affecting the study area, the azimuths may not be fully restored to their original orientations.

To arrive at the paleocurrent azimuth of each measured cross-bed, ninety degrees was either added to, or subtracted from, the corrected foreset strike (tables 1 and 2), as the dip direction of each cross-bed required. The widely-used (Greene, 1984; Bowden, 1977; Miall, 1974) vector method of Curray (1956) was employed to further process this paleocurrent data.

Each cross-bed set contributes a single paleocurrent azimuth to the calculation of an overall paleocurrent rose, so the vector mean azimuth of each set which yielded more than one measurement was calculated. The vector mean is a measure of the central tendency of the input azimuths, and was calculated with the following equation (Curray, 1956):

$$0 = \arctan \Sigma n \text{ Sin } 0_i / \Sigma n \text{ Cos } 0_i$$

where 0 is the vector mean azimuith; n equals the number of observations; and 0_i equals the azimuth of the foreset dip direction (Curray, 1956).

Because the arctan function returns the vector mean azimuth (0) as a negative number in the fourth quadrant, or a positive number in the first quadrant: if the vector resulting from this equation should be in the SW or SE quadrants, add 180 to 0; if the resultant vector mean azimuth belongs in the NW quadrant, add 360 to 0; if it belongs in the NE quadrant, the value of 0 is correct unmodified.

Since vectors possess magnitude as well as direction, it is necessary to calculate the vector magnitude of each resultant vector. The vector magnitude is a measure of the concentration of the azimuths as they are arranged about the vector mean, and is calculated using the following equation (Curray, 1956):

$$r = ((\Sigma n \sin 0_i)^2 + (\Sigma n \text{ Cos } 0_i)^2)^{1/2}$$

where r is the vector magnitude.

Since the vector magnitude is a unitless number, the significance of which is difficult to visualize, the consistency ratio, or vector magnitude in terms of percent, is calculated by the following expression (Curray, 1956):

$$L = r / \Sigma n \times 100$$

where L is the consistency ratio; r equals the vector magnitude; and n equals the number of observations.

The consistency ratio (0-100%) is a measure of dispersion of measurements about the vector mean. For example: when L=100%, all the data are the same; when L=0%, the data is uniformly distributed about the vector mean. Tables 1 and 2 lists 0, and L for each cross-bed set.

In constructing a rose diagram, each cross-bed sets was represented by its vector mean azimuth, and a frequency distribution was constructed. The azimuths were grouped into twenty degree intervals (0-19, 20-29, etc.), and the number in each interval was converted to a percentage by dividing it by the total number of azimuths. The rose diagram was constructed on polar coordinate paper with the line segment radius directly proportional to the square-root of the interval percentage (Greene, 1984; Miall, 1974), so that the value plotted is directly proportional to the interval area rather than to the segment length.

APPENDIX C: LOCATION OF MEASURED SECTIONS

Legend

Em - Empire Formation

SR - St. Regis Fm. and correlative Grinnell Fm.

uR - upper Revett Fm. and correlative Grinnell Fm.

mR - middle Revett Fm. and correlative Grinnell Fm.

lR - lower Revett Fm. and correlative Grinnell Fm.

Bu - Burke Fm. and correlative Grinnell Fm.

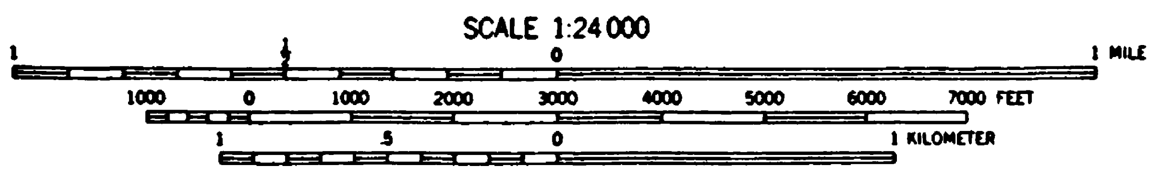
B - Unit B

A - Unit A

uP - upper unit of Prichard Fm.

lP - lower unit of Prichard Fm.

① - Subsection number



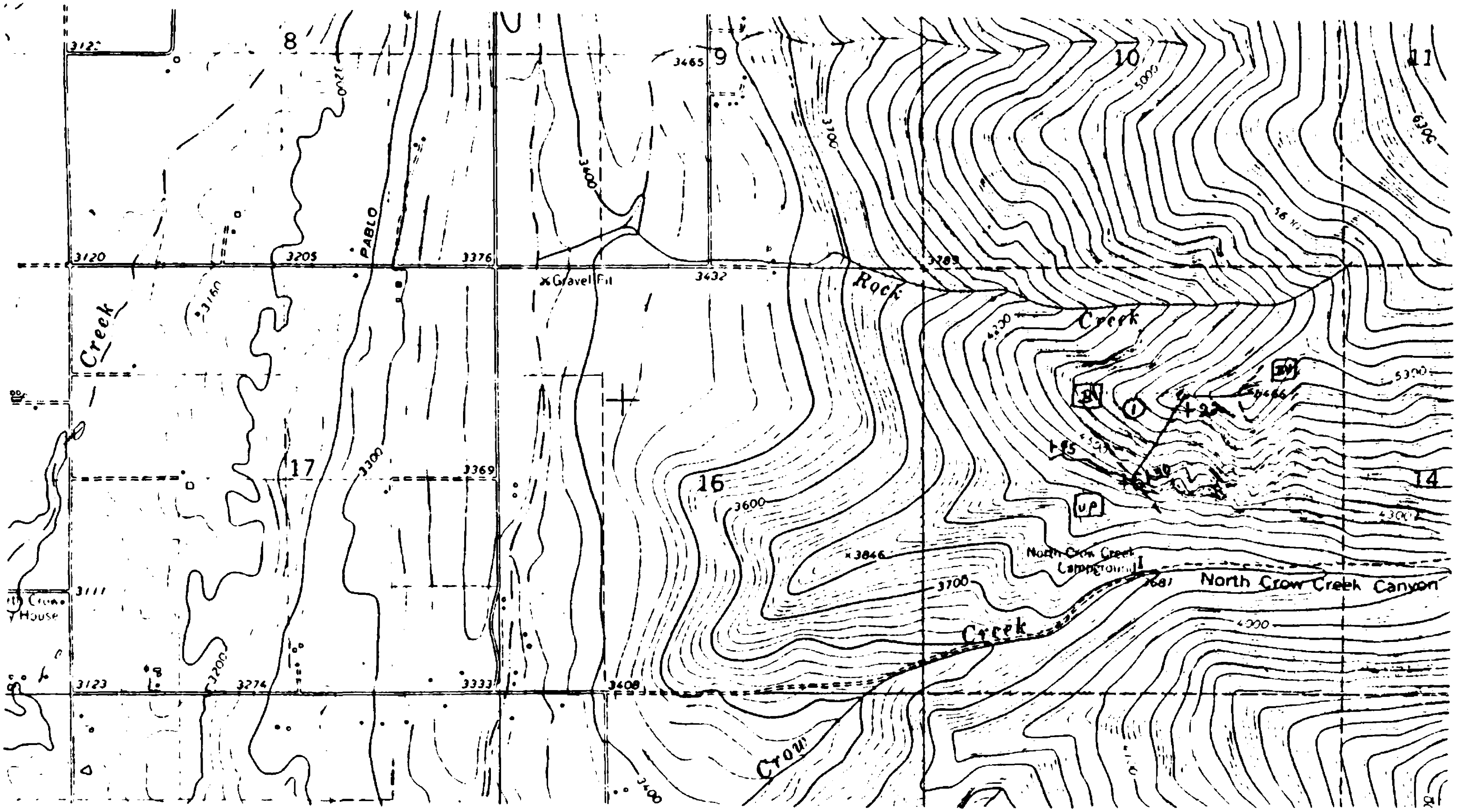


Figure C-1. North Crow Creek; Ronan Quadrangle.

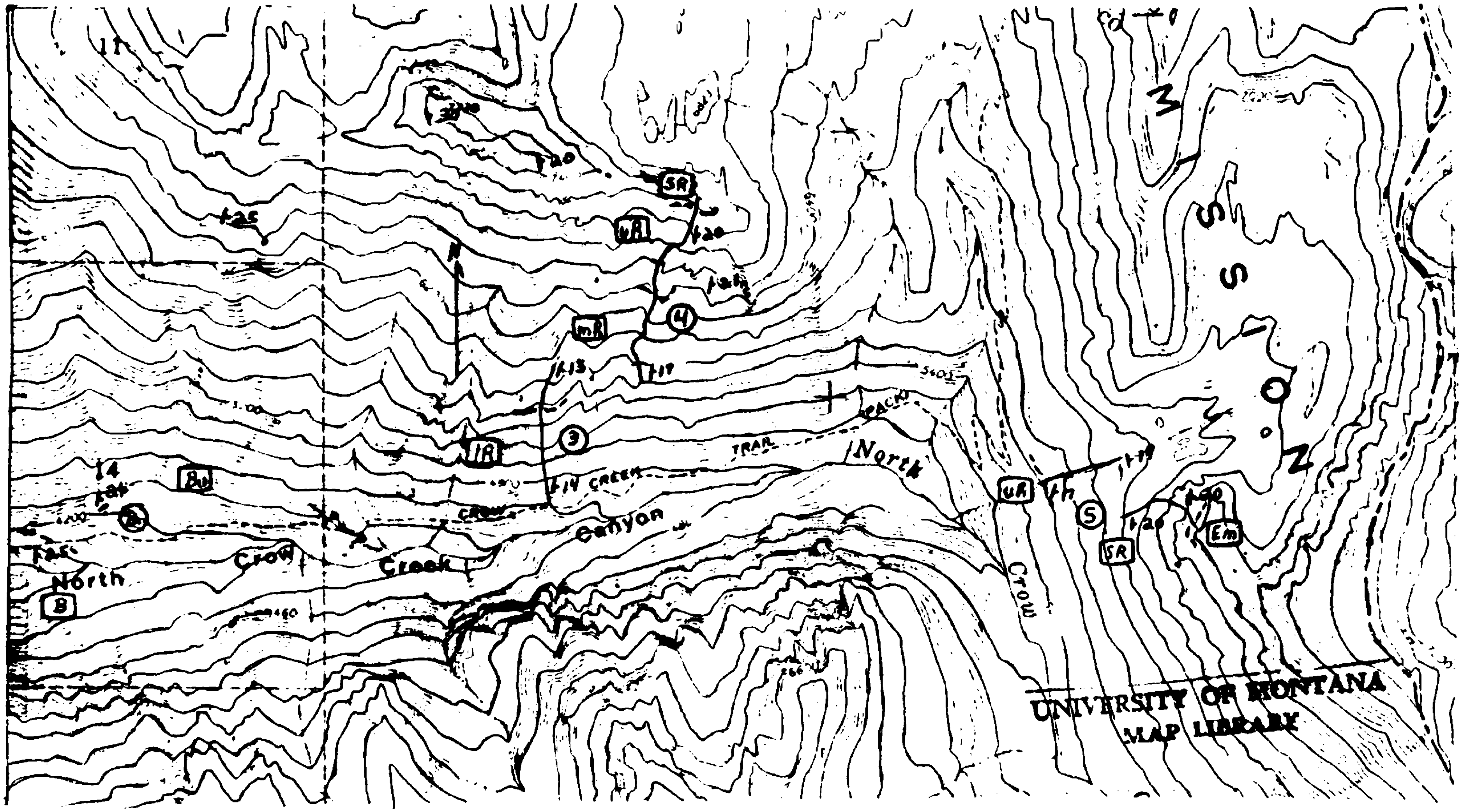


Figure C-2. North Crow Creek; Piper-Crow Pass Quadrangle.

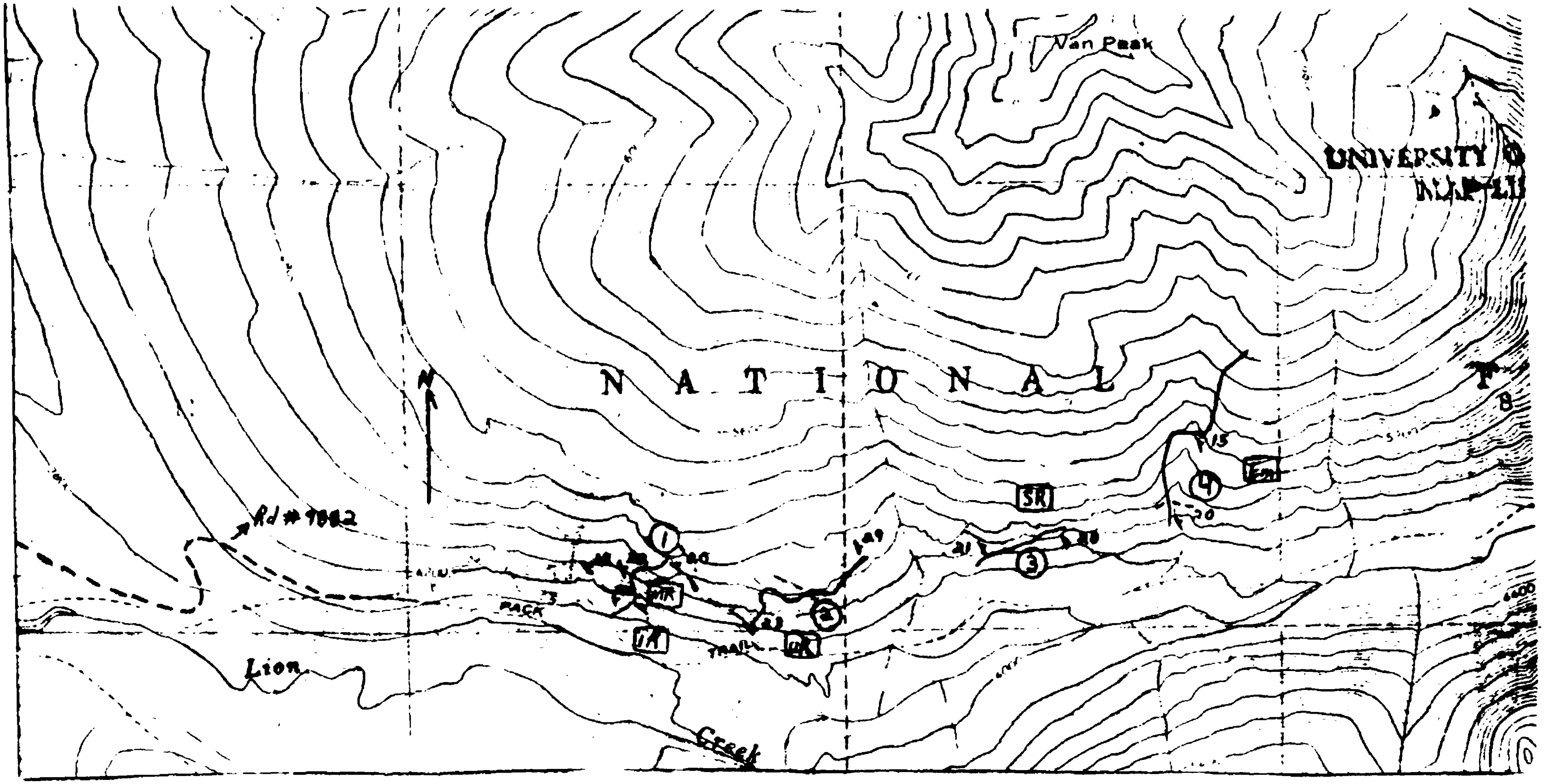


Figure C-3. Lion Creek; Swan Peak Quadrangle.

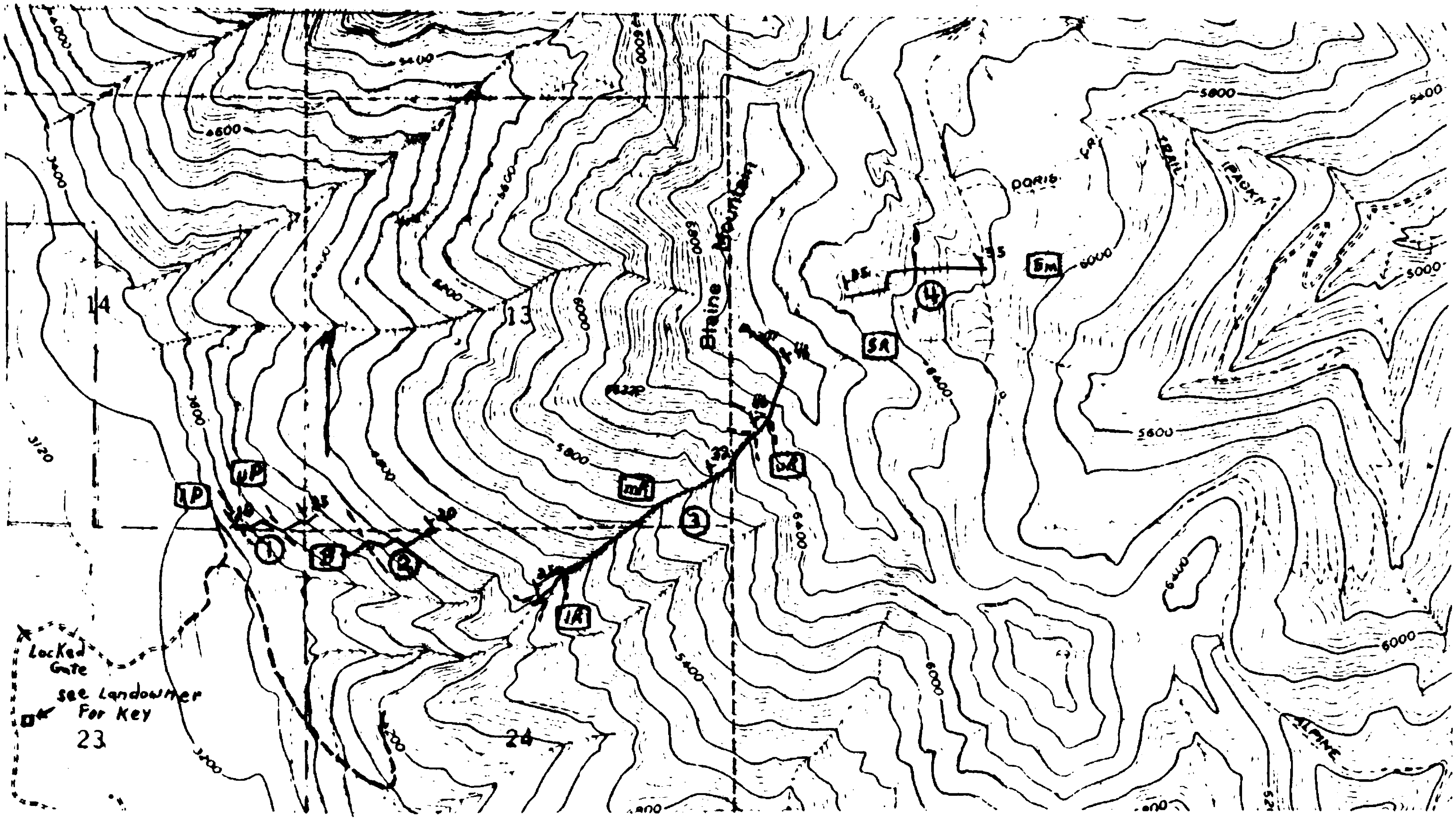


Figure C-4. Blaine Mountain; Blaine Mtn. Quadrangle.

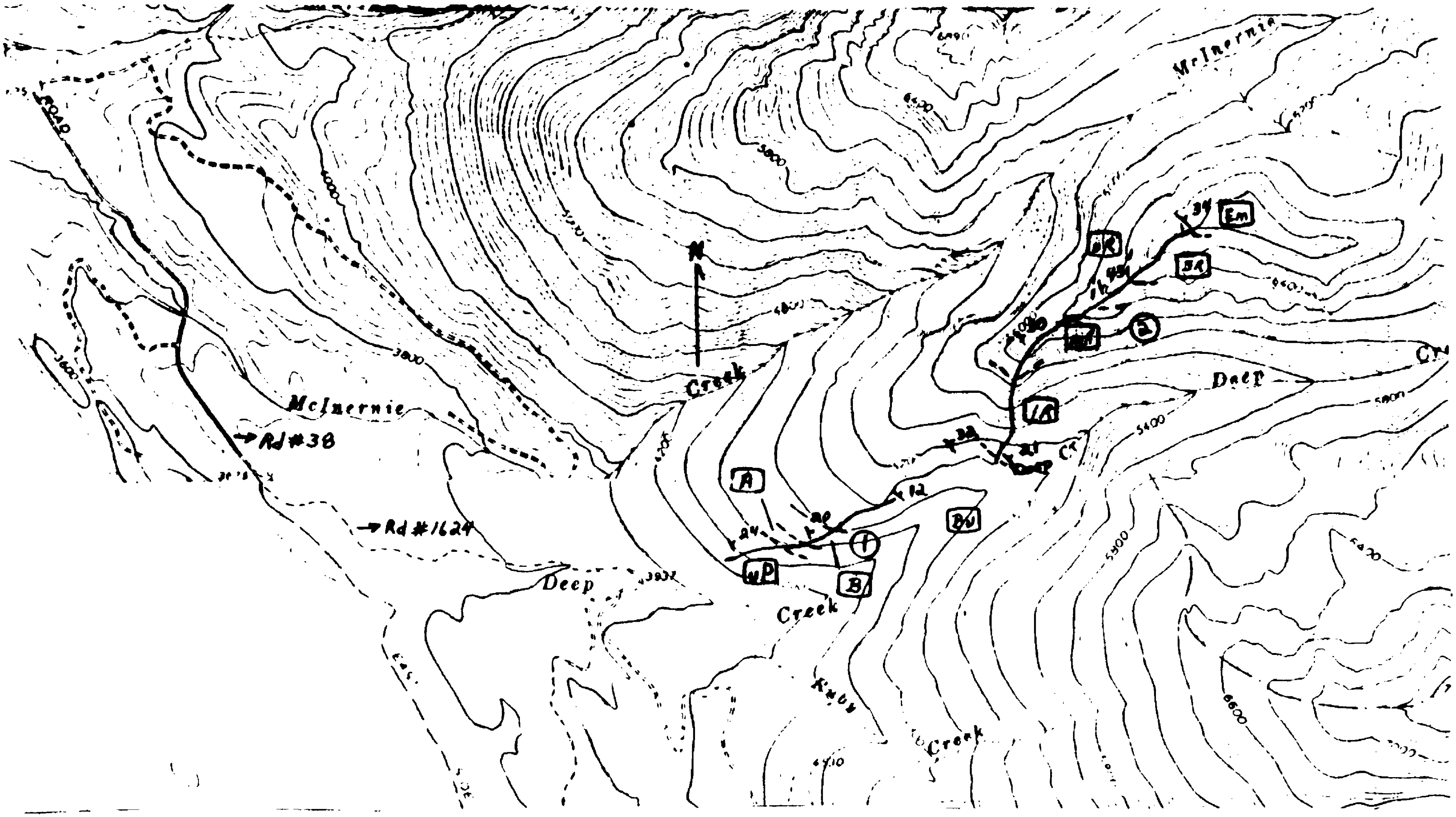


Figure C-5. Deep Creek; Pinnacle, Mt. Grant, Felix Peak, and Pioneer Ridge.

APPENDIX D

APPROXIMATION OF THE THICKNESS OF THE NORTH CROW CREEK COVERED INTERVAL

A large forested area with little outcrop covers most of the Burke Formation and the lower part of the lower Revett Formation at North Crow Creek. The thickness of this covered interval was approximated using the construction shown in figure 40.

First, a topographic cross-section was constructed along line A-A'. Next, the bases of sub-sections 2 and 3 were located on the cross section. Lines inclined at the dip of bedding were drawn from these two points, and from the top of sub-section 1. The thickness between the top of sub-section 1 and the base of sub-section 2 is approximately 500 feet, and the gap between the bases of sub-sections 2 and 3 is approximately 1500 feet. Covered intervals of these thicknesses were installed in the North Crow Creek stratigraphic section.

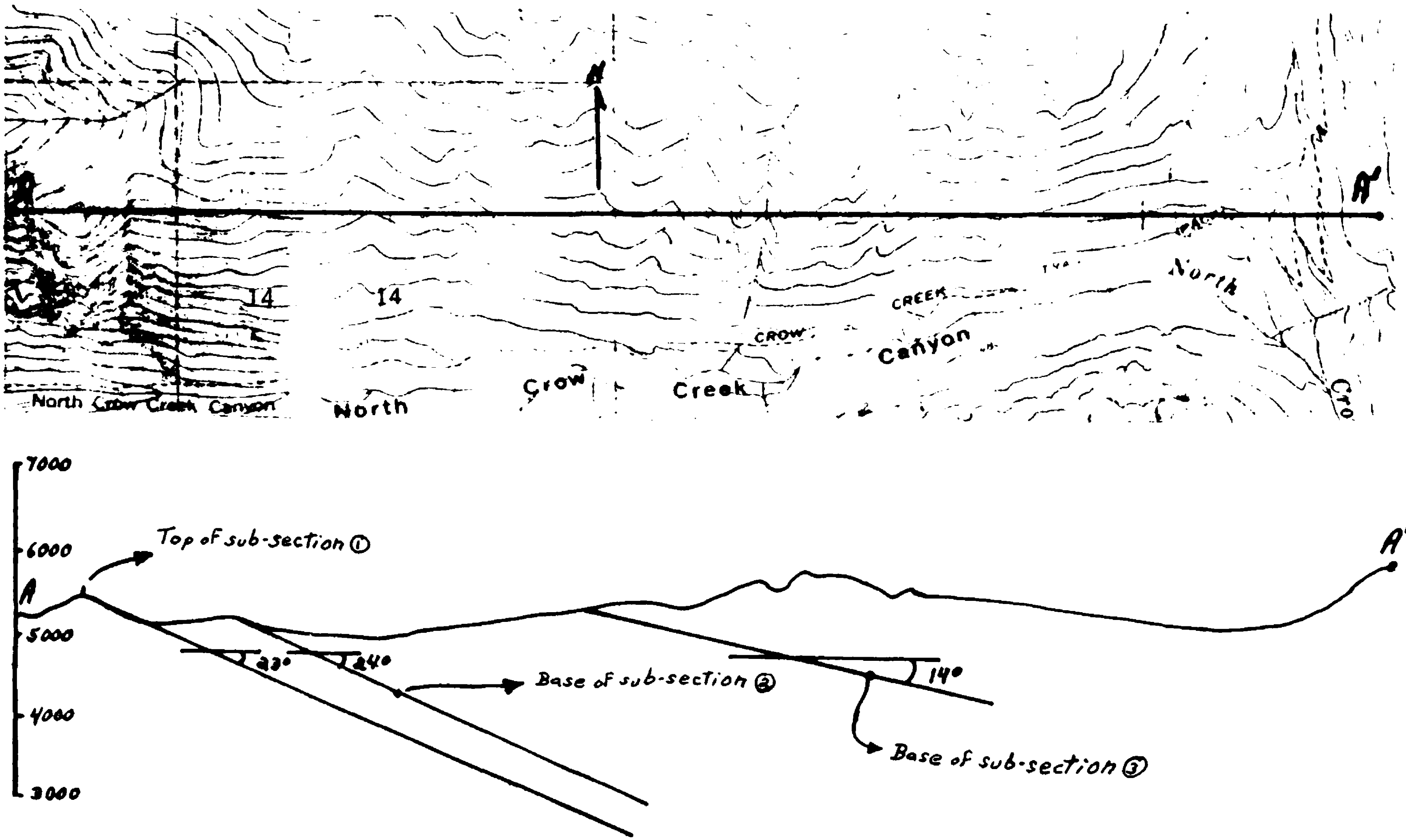
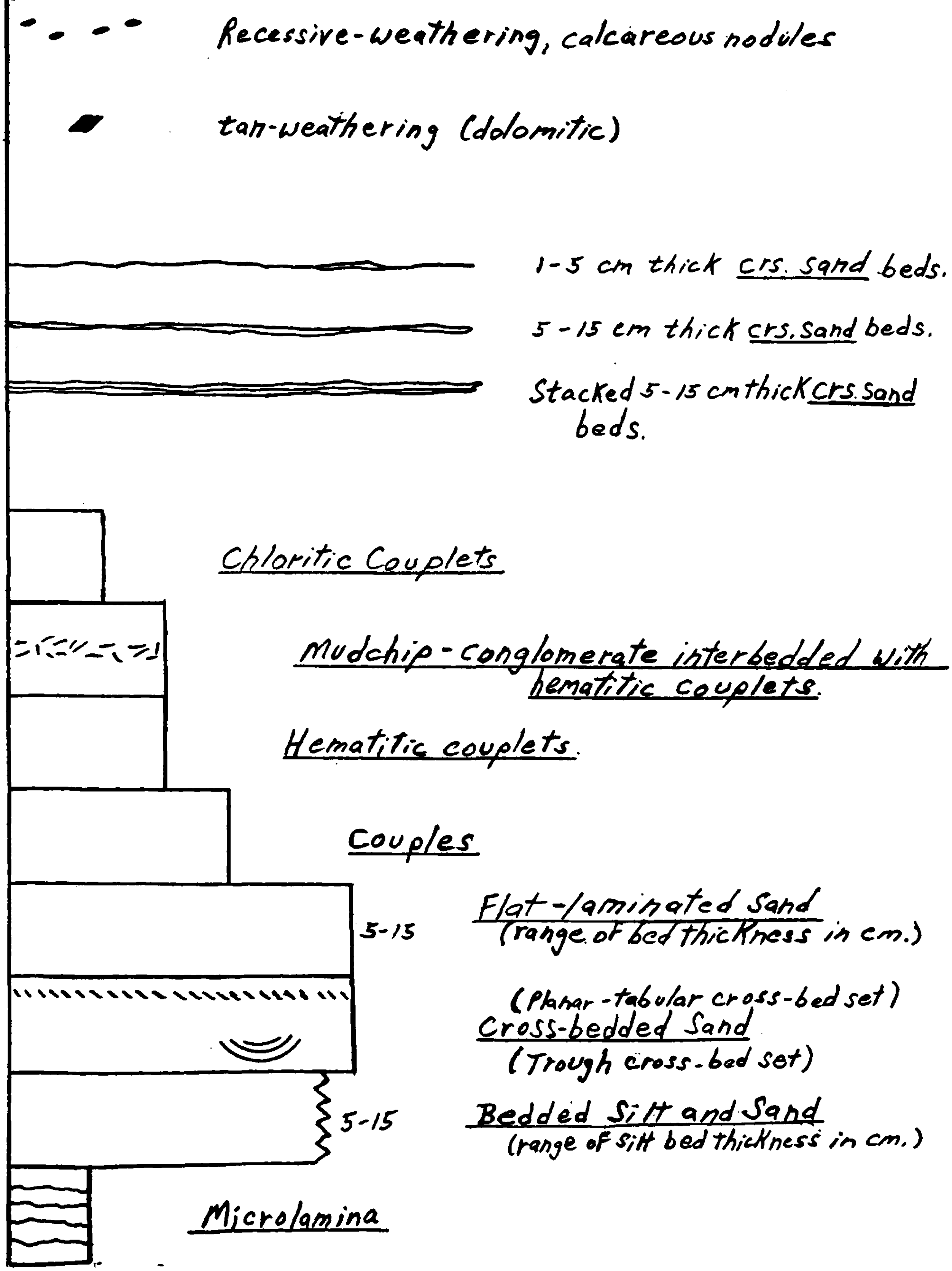

A
A'
North Crow Creek Canyon
14
14
North
Crow
Creek
CROW
CREEK
Canyon
North
7000
6000
5000
4000
3000
Top of sub-section ①
23°
24°
Base of sub-section ②
14°
Base of sub-section ③

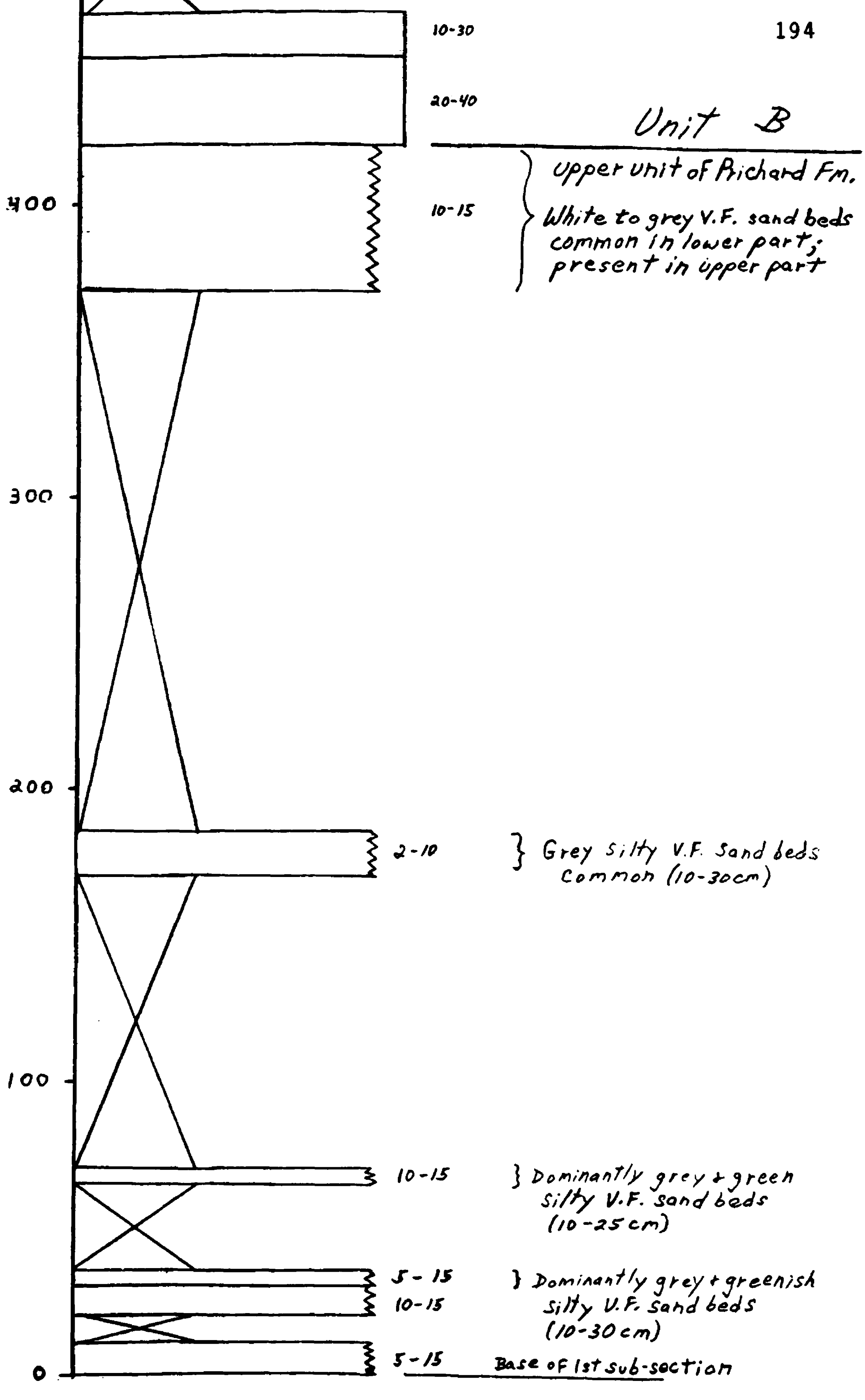
Figure D-1.

Legend

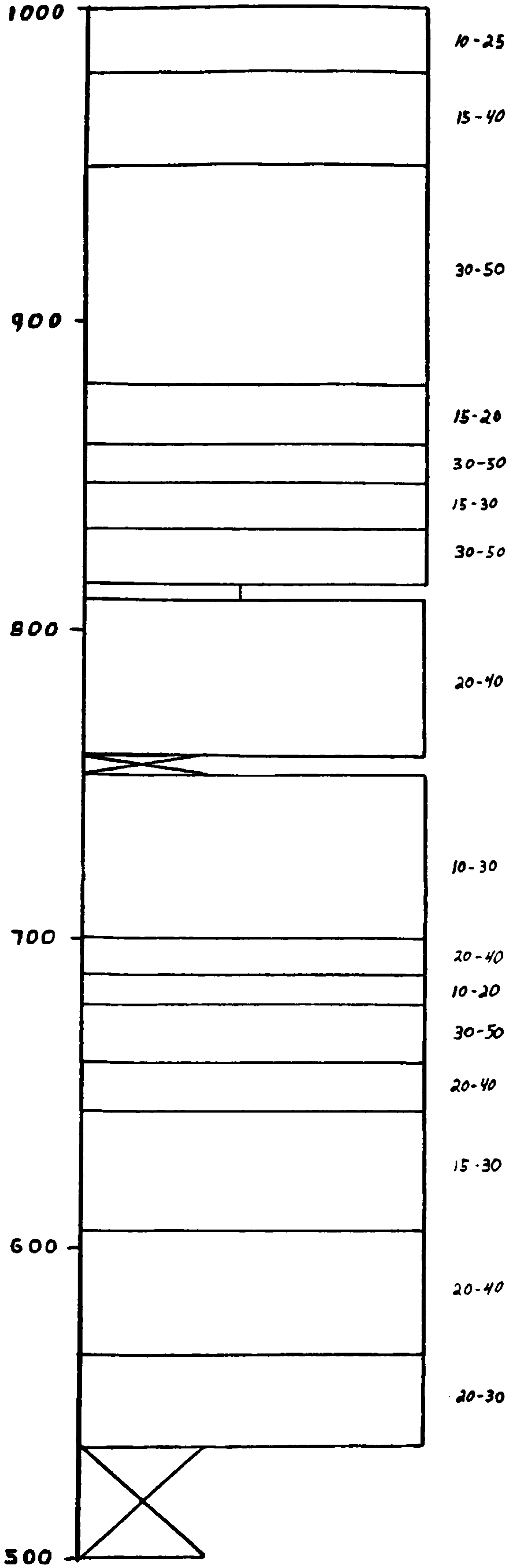
Scale: 1 inch = 50 Feet



North Crow Creek

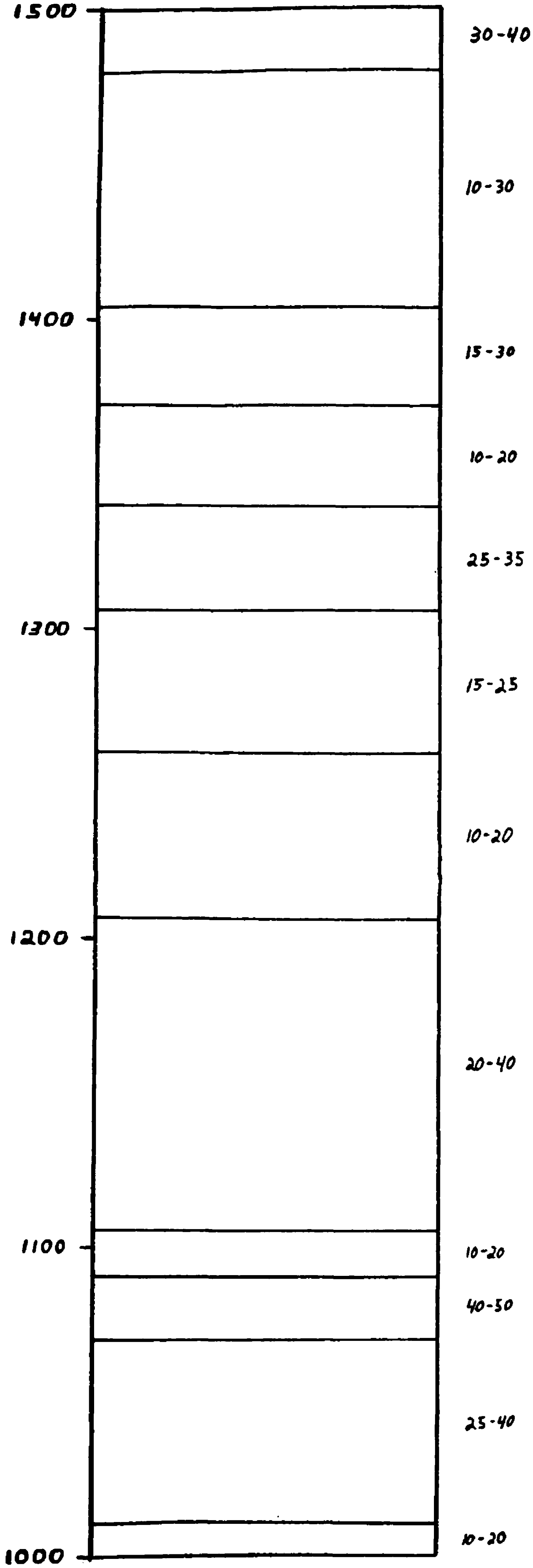


NCC

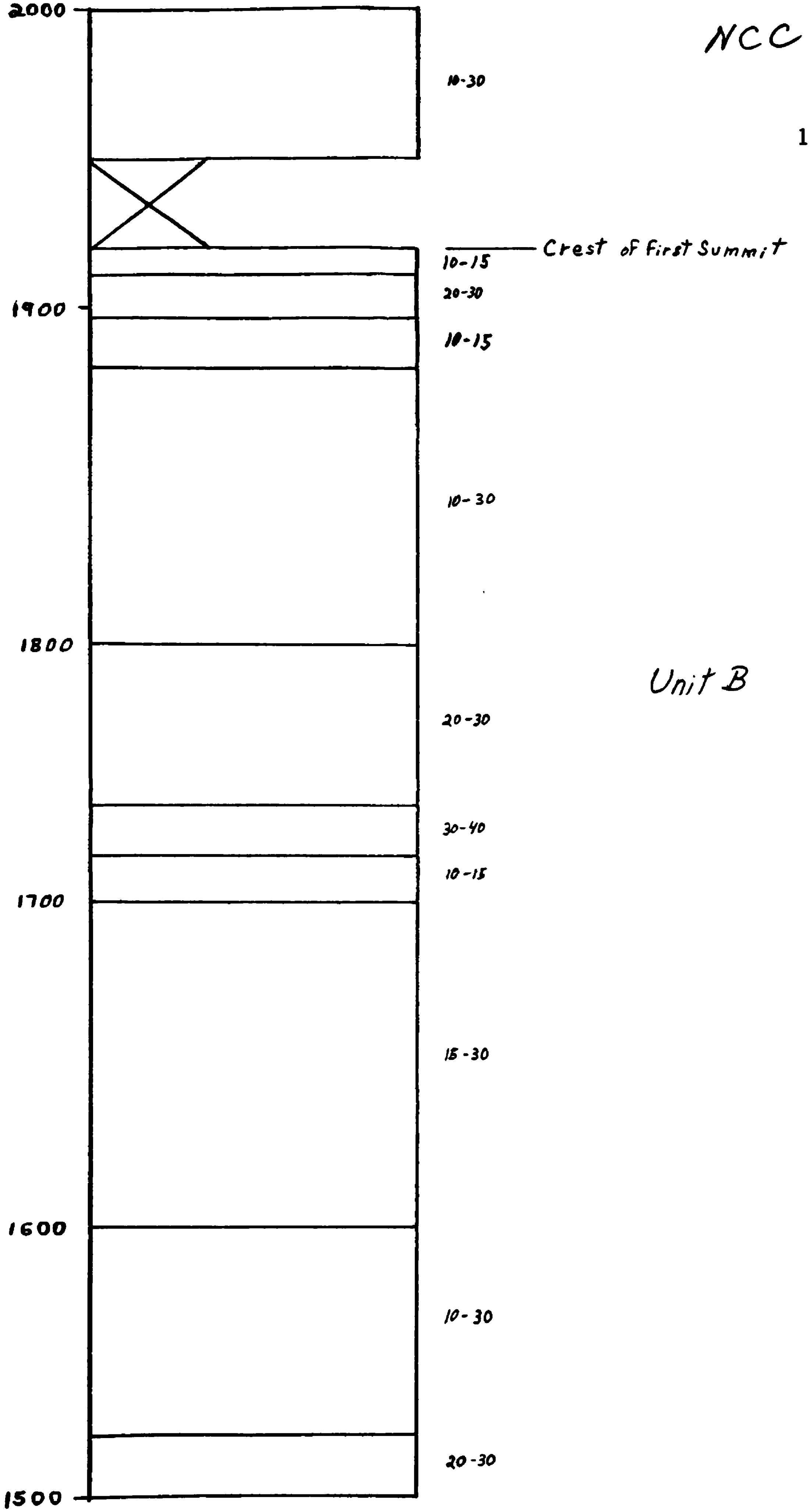

1000
10-25
15-40
30-50
900
15-20
30-50
15-30
30-50
800
20-40
10-30
700
20-40
10-20
30-50
20-40
15-30
600
20-40
20-30
500

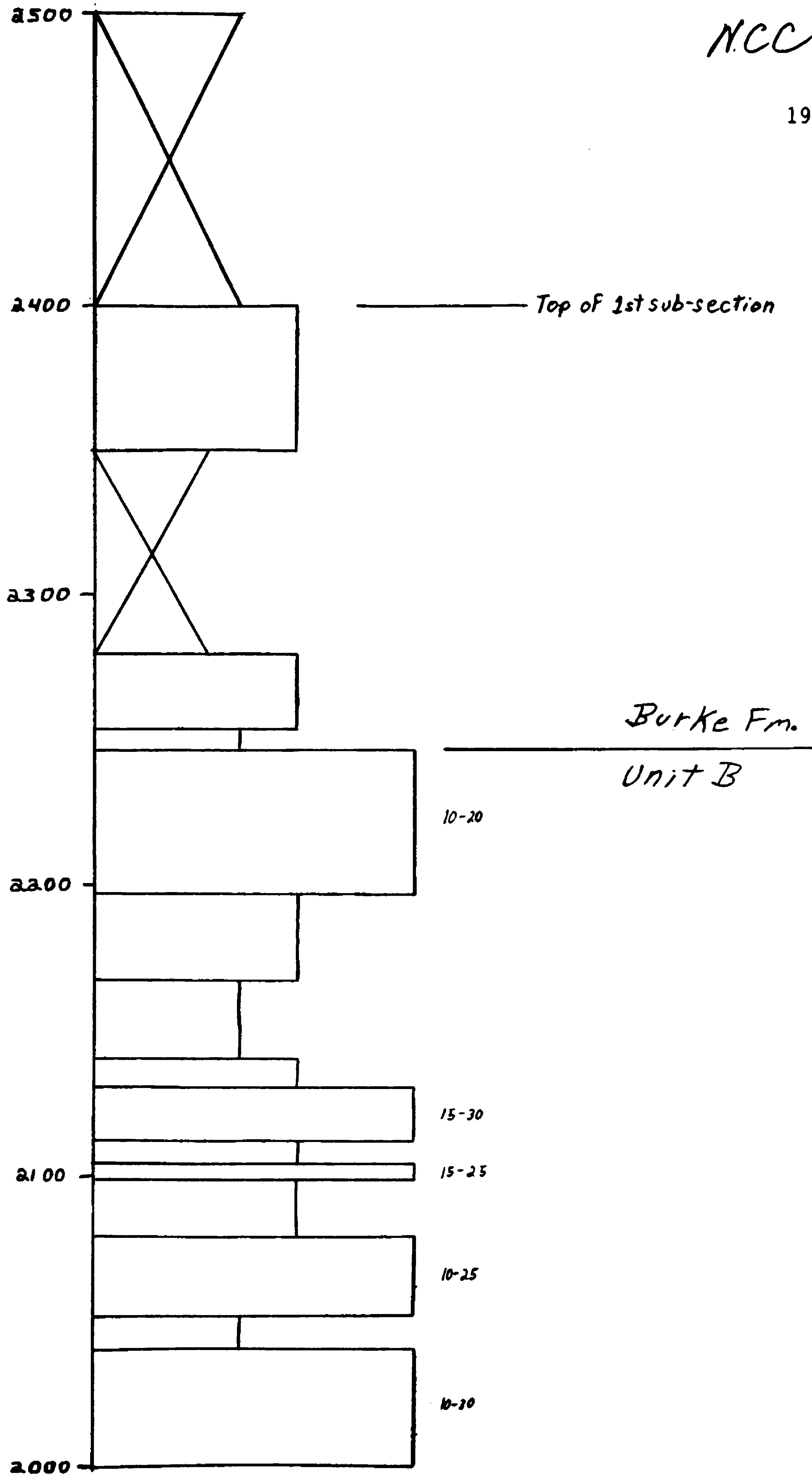
Unit B

NCC

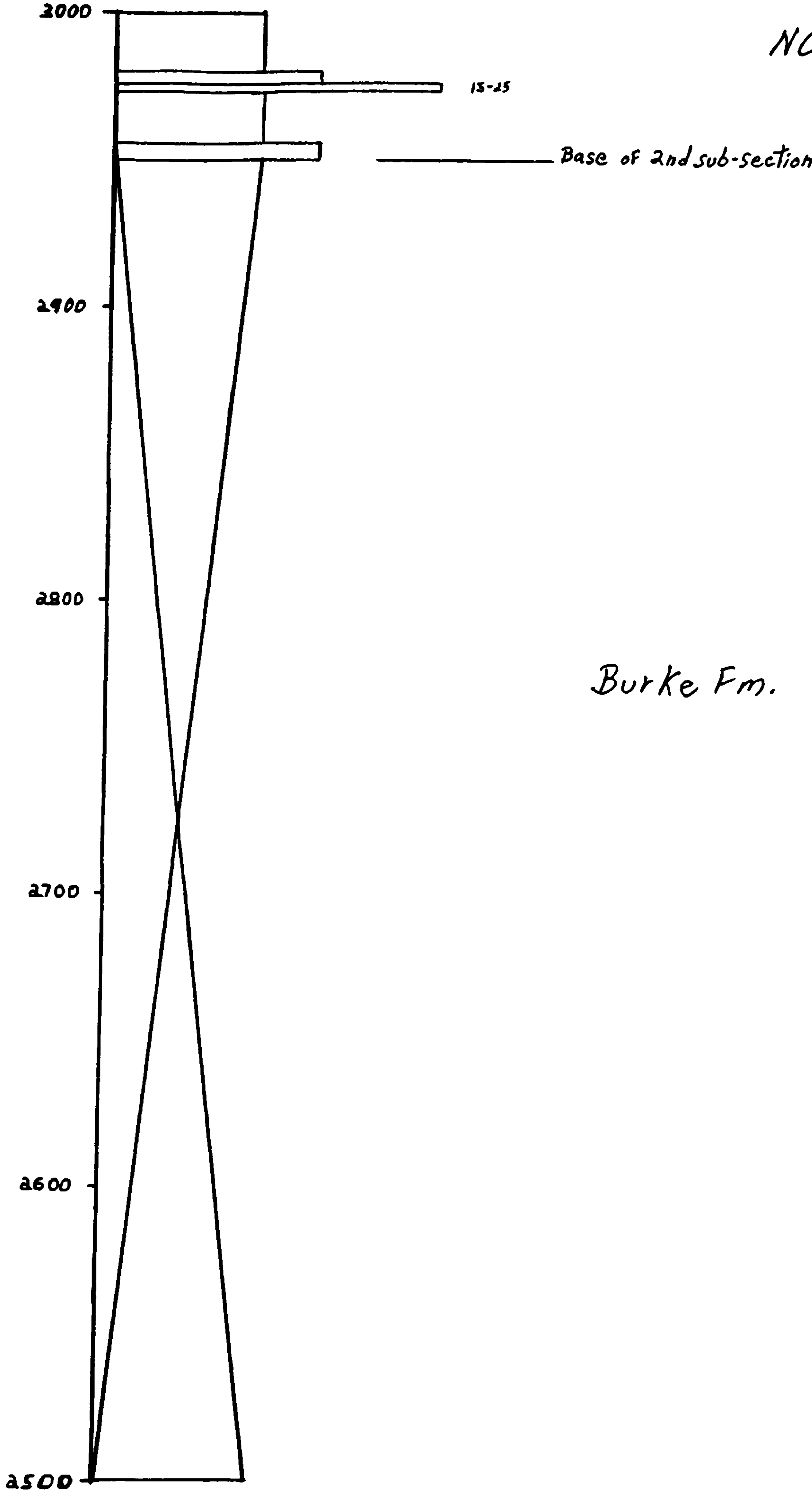

1500
1400
1300
1200
1100
1000
30-40
10-30
15-30
10-20
25-35
15-25
10-20
20-40
10-20
40-50
25-40
10-20

Unit B

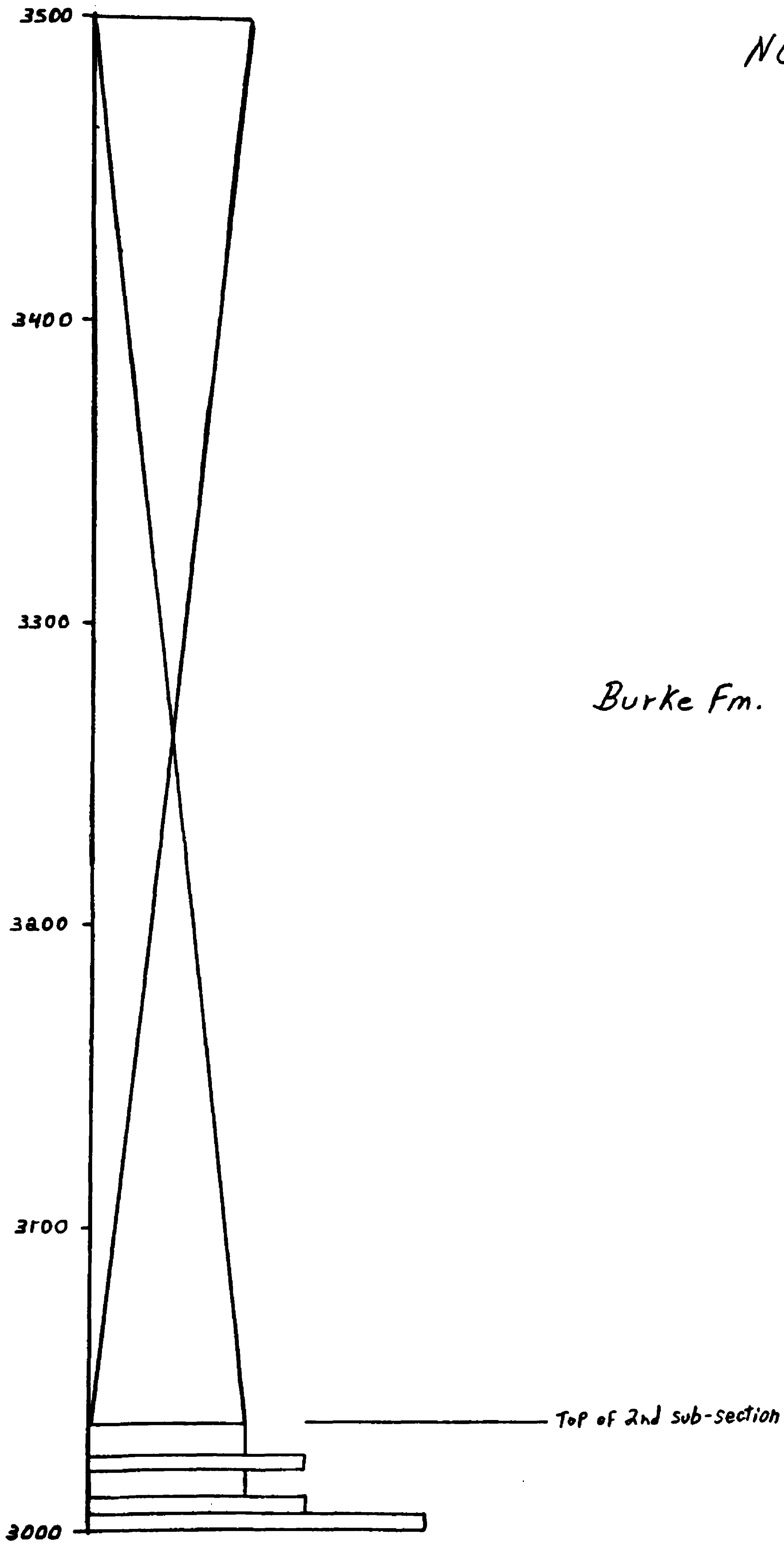

NCC
2000
10-30
Crest of First Summit
10-15
20-30
1900
10-15
10-30
1800
Unit B
20-30
30-40
10-15
1700
15-30
1600
10-30
20-30
1500

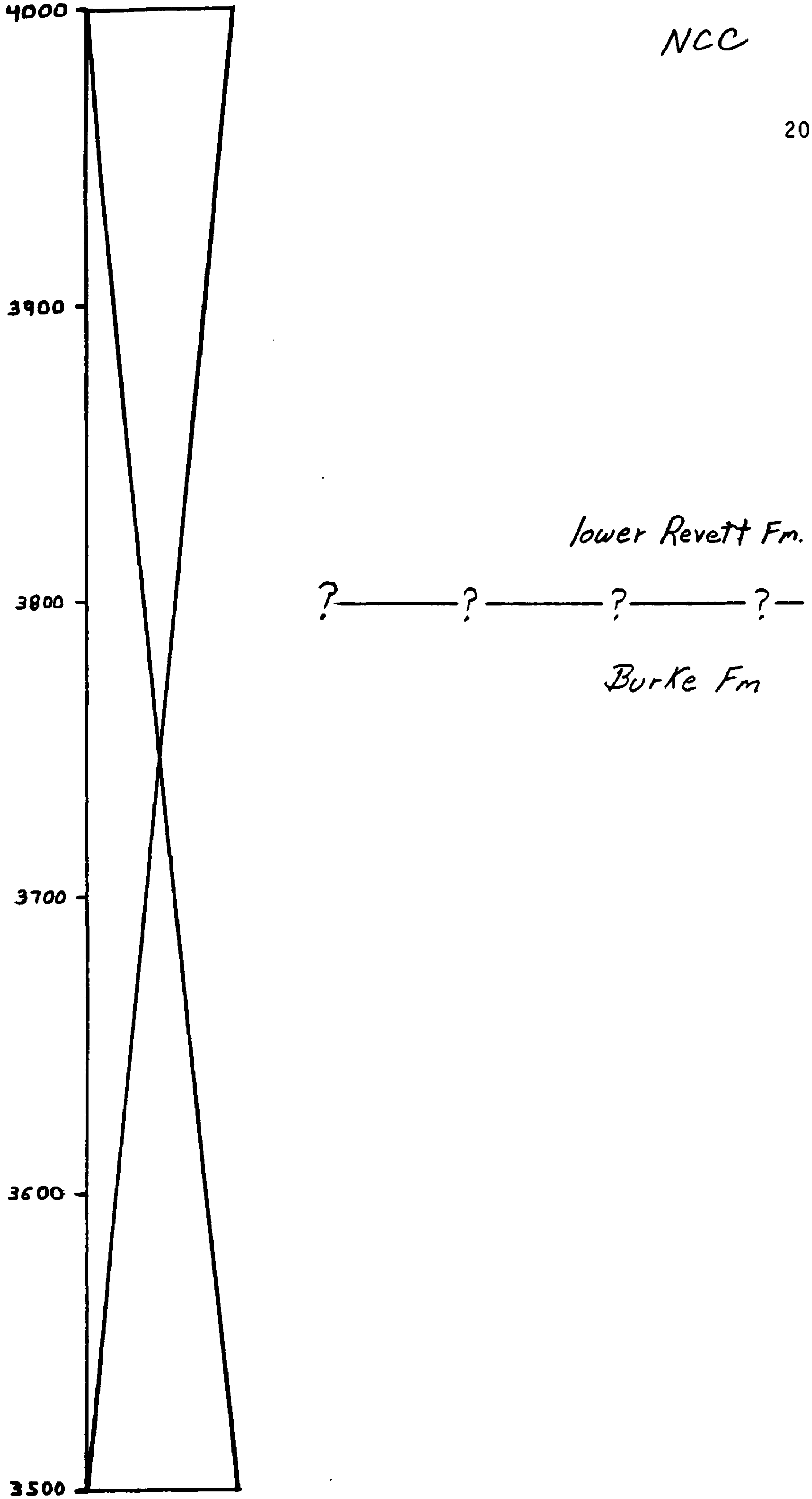

N.CC
2500
2400
Top of 1st sub-section
2300
Burke Fm.
Unit B
10-20
2200
15-30
2100
15-25
10-25
10-20
2000

NCC


2000
15-25
Base of 2nd sub-section
2900
2800
Burke Fm.
2700
2600
2500

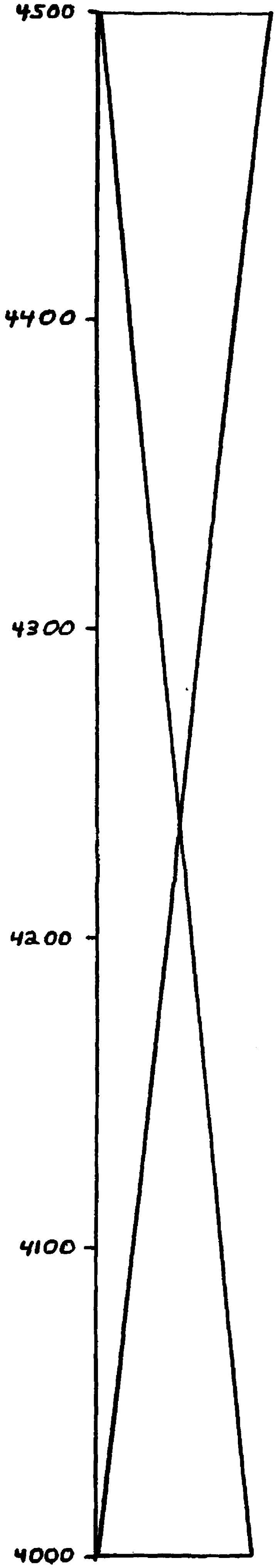
NCC



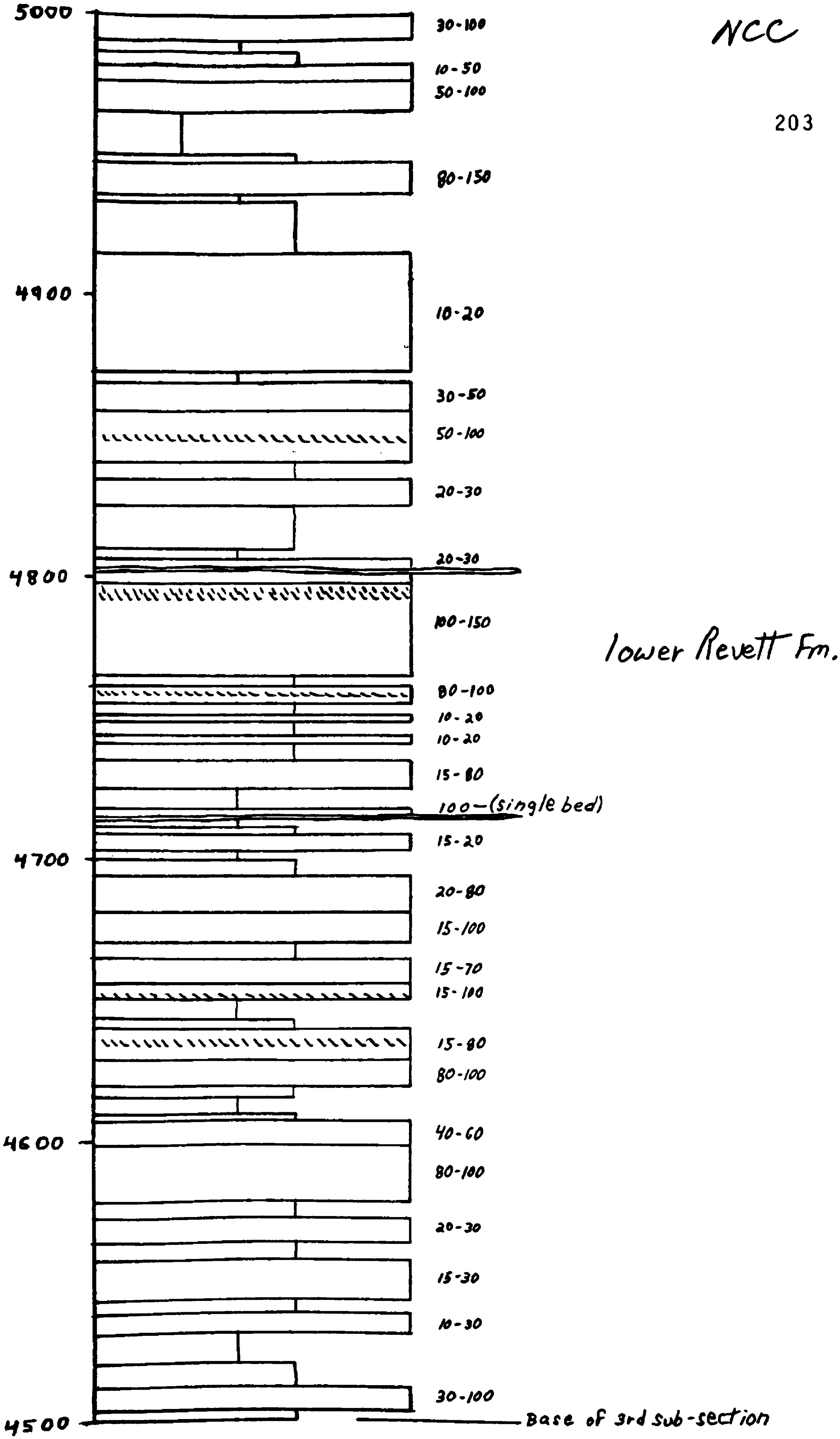

NCC
4000
3900
3800
3700
3600
3500
lower Revett Fm.
?——?——?——?—
Burke Fm

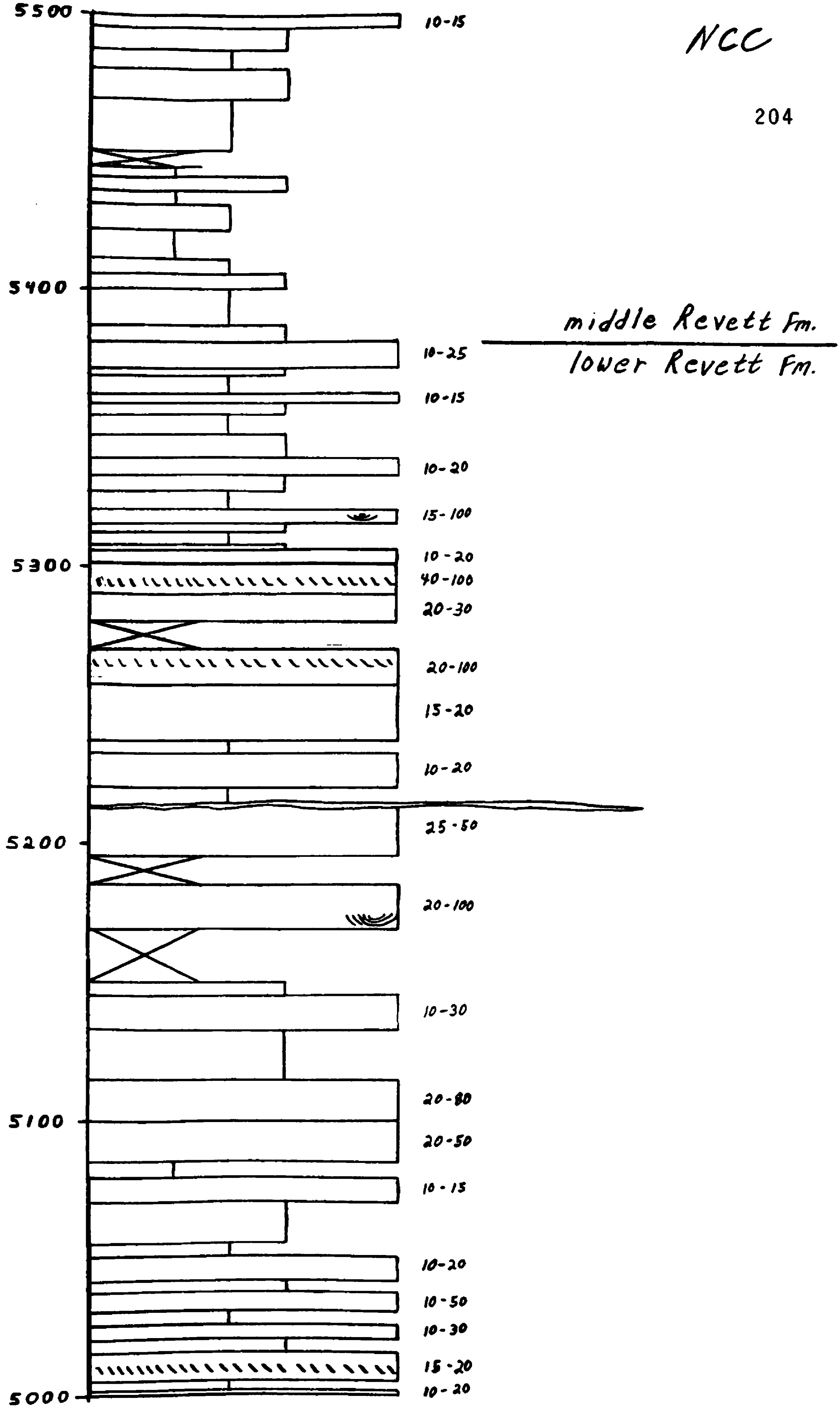
NCC

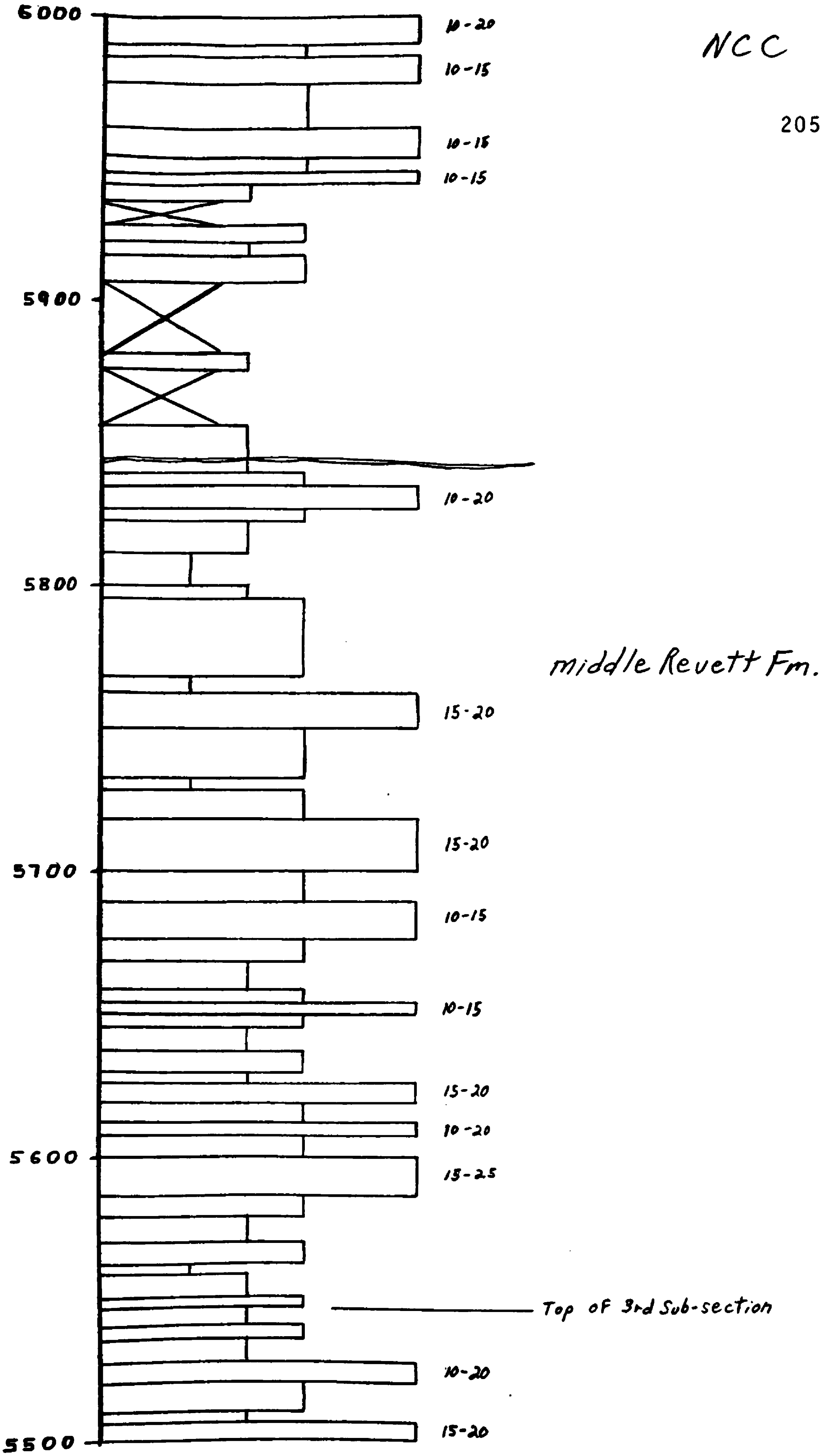
202

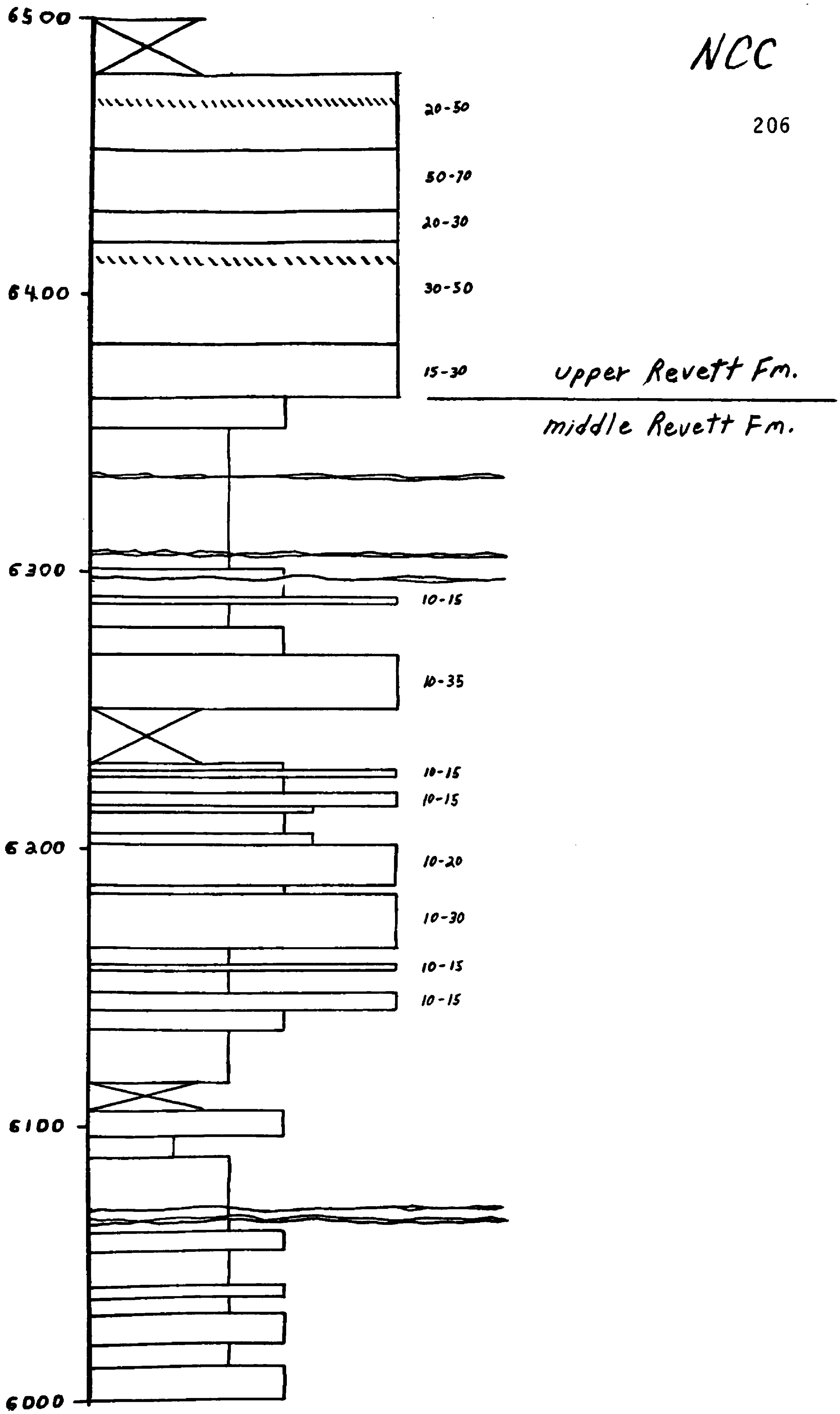


lower Revett Fm

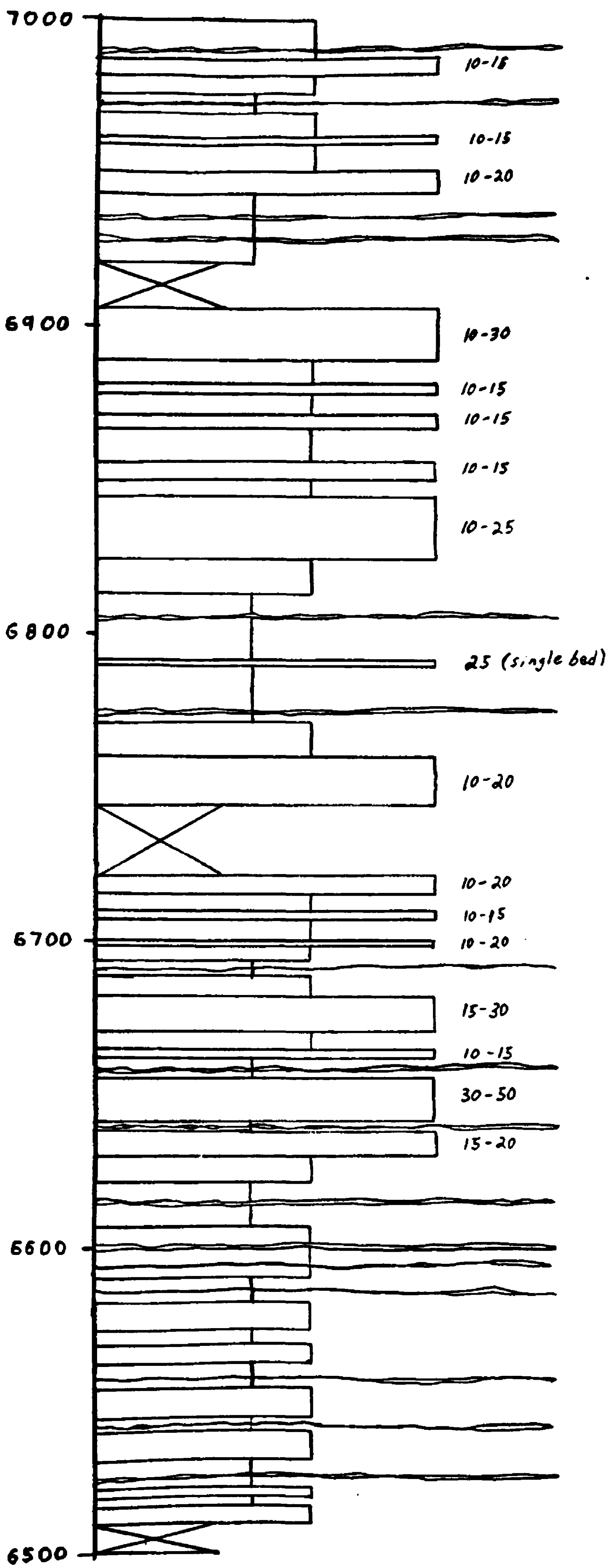

NCC
5000
30-100
10-50
50-100
80-150
4900
10-20
30-50
50-100
20-30
20-30
4800
100-150
lower Revett Fm.
80-100
10-20
10-20
15-80
100—(single bed)
15-20
4700
20-80
15-100
15-70
15-100
15-80
80-100
40-60
4600
80-100
20-30
15-30
10-30
30-100
4500
Base of 3rd sub-section


NCC
5500
10-15
5400
middle Revett Fm.
lower Revett Fm.
10-25
10-15
10-20
15-100
10-20
5300
40-100
20-30
20-100
15-20
10-20
25-50
5200
20-100
10-30
20-80
5100
20-50
10-15
10-20
10-50
10-30
15-20
10-20
5000

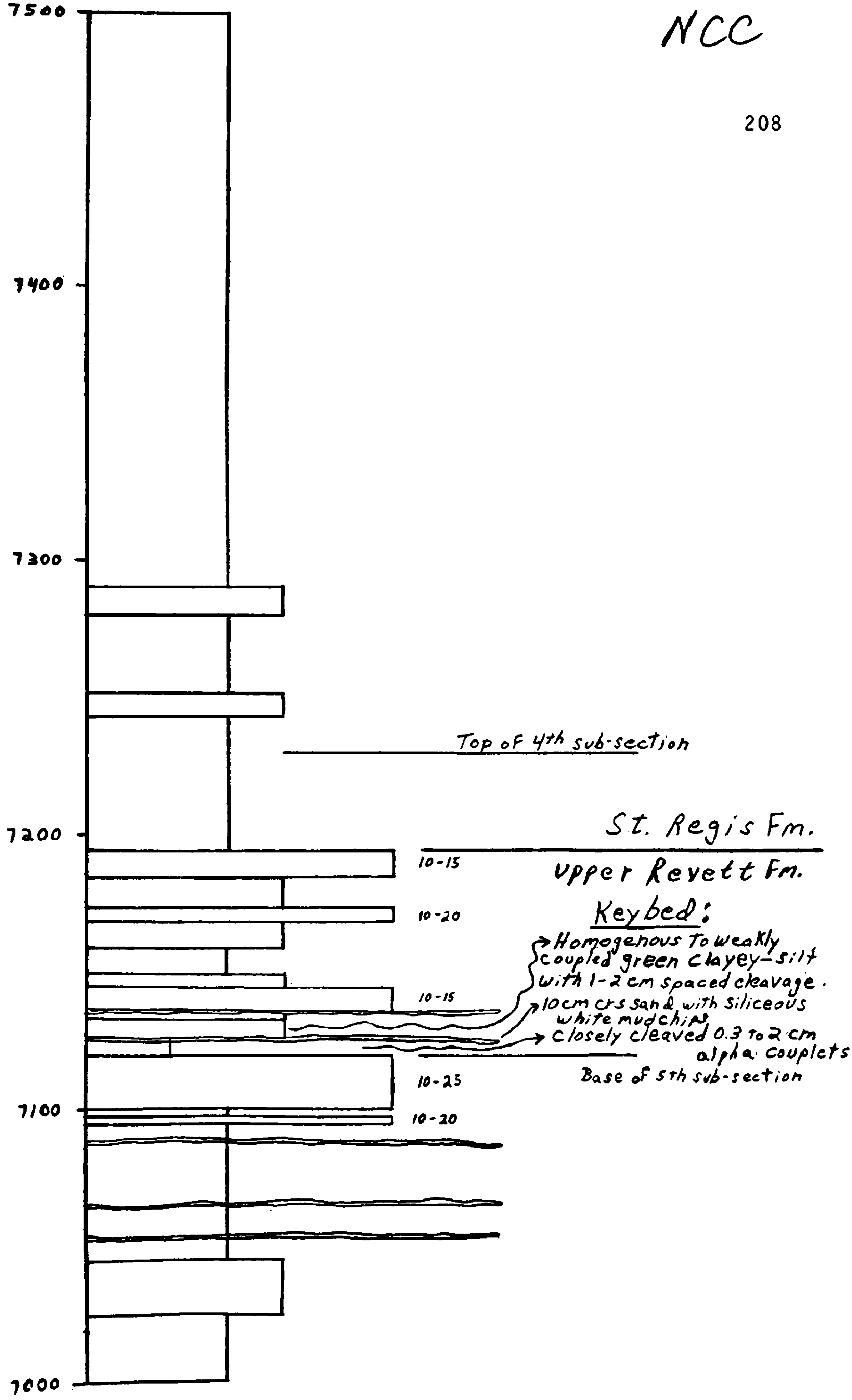

NCC
6000
10-20
10-15
10-15
10-15
5900
10-20
5800
middle Revett Fm.
15-20
15-20
5700
10-15
10-15
15-20
10-20
5600
15-25
Top of 3rd Sub-section
10-20
15-20
5500

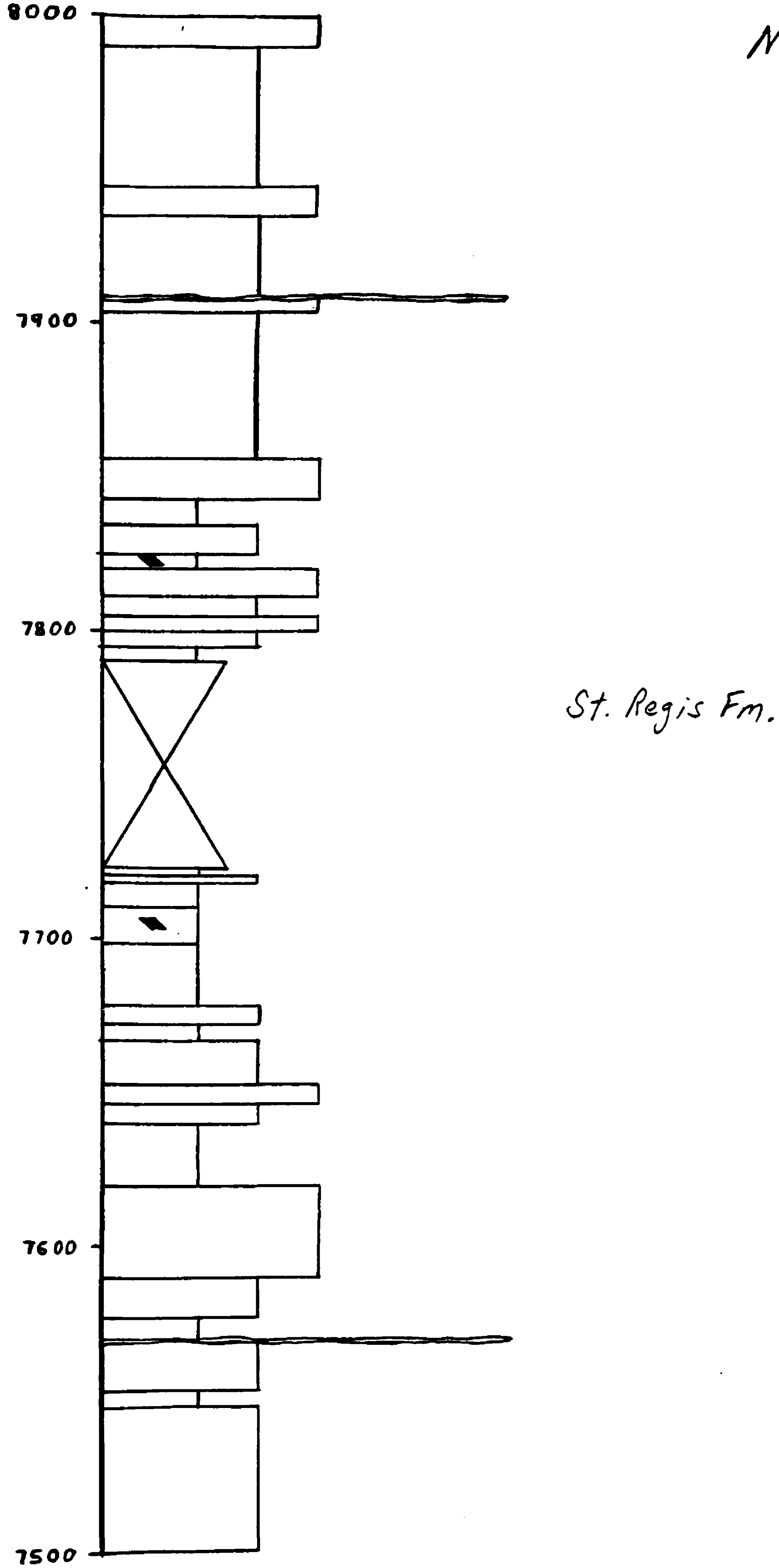

NCC
206
6500
20-50
50-70
20-30
6400
30-50
15-30
upper Revett Fm.
middle Revett Fm.
6300
10-15
10-35
10-15
10-15
6200
10-20
10-30
10-15
10-15
6100
6000

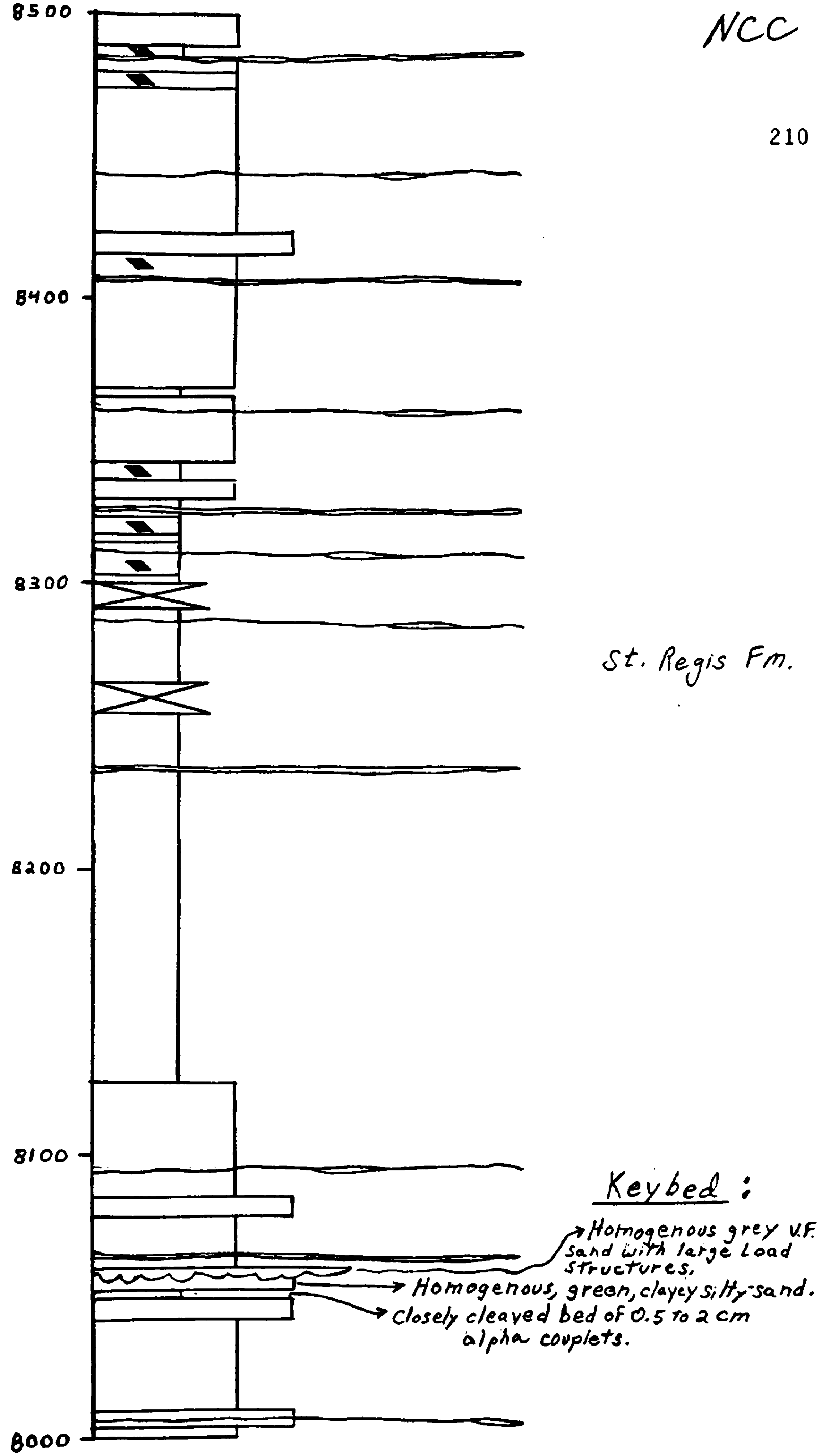
NCC

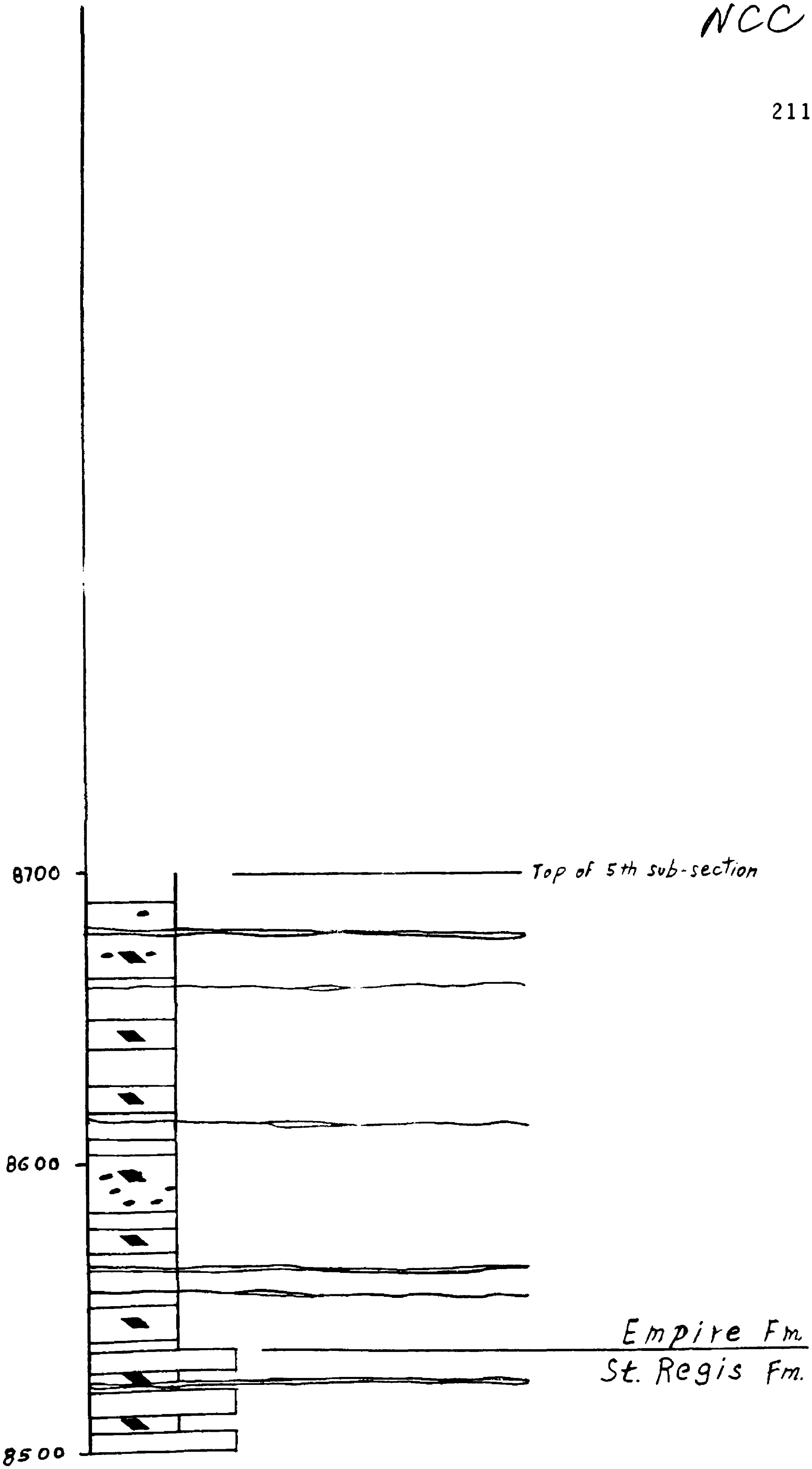


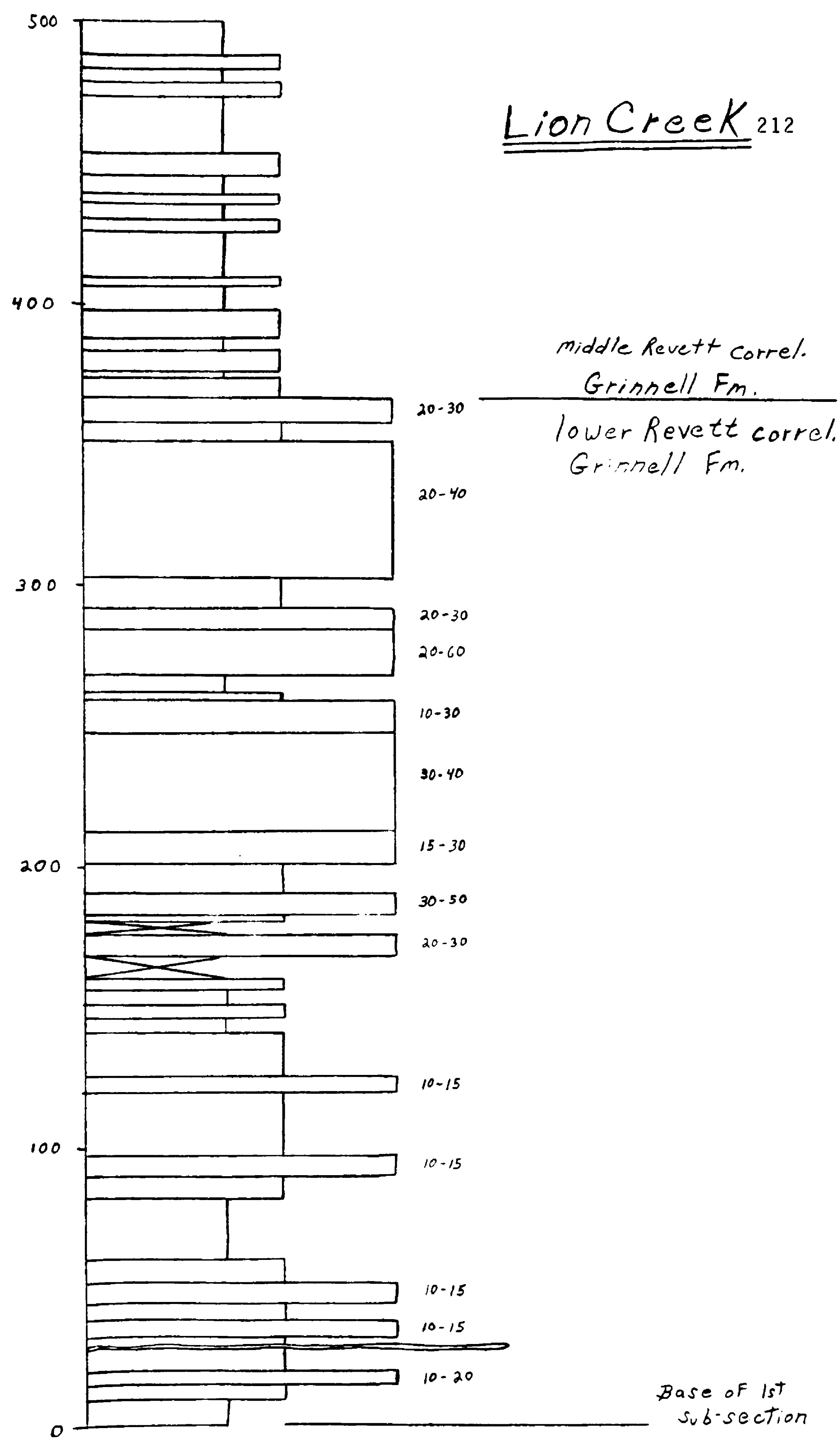
Upper Revett Fm.


7500
7400
7300
Top of 4th sub-section
7200
St. Regis Fm.
upper Revett Fm.
Keybed:
Homogenous to weakly coupled green clayey-silt with 1-2 cm spaced cleavage.
10cm crs sand with siliceous white mudchips
Closely cleaved 0.3 to 2 cm alpha couplets
Base of 5th sub-section
10-15
10-20
10-15
10-25
7100
10-20
7000

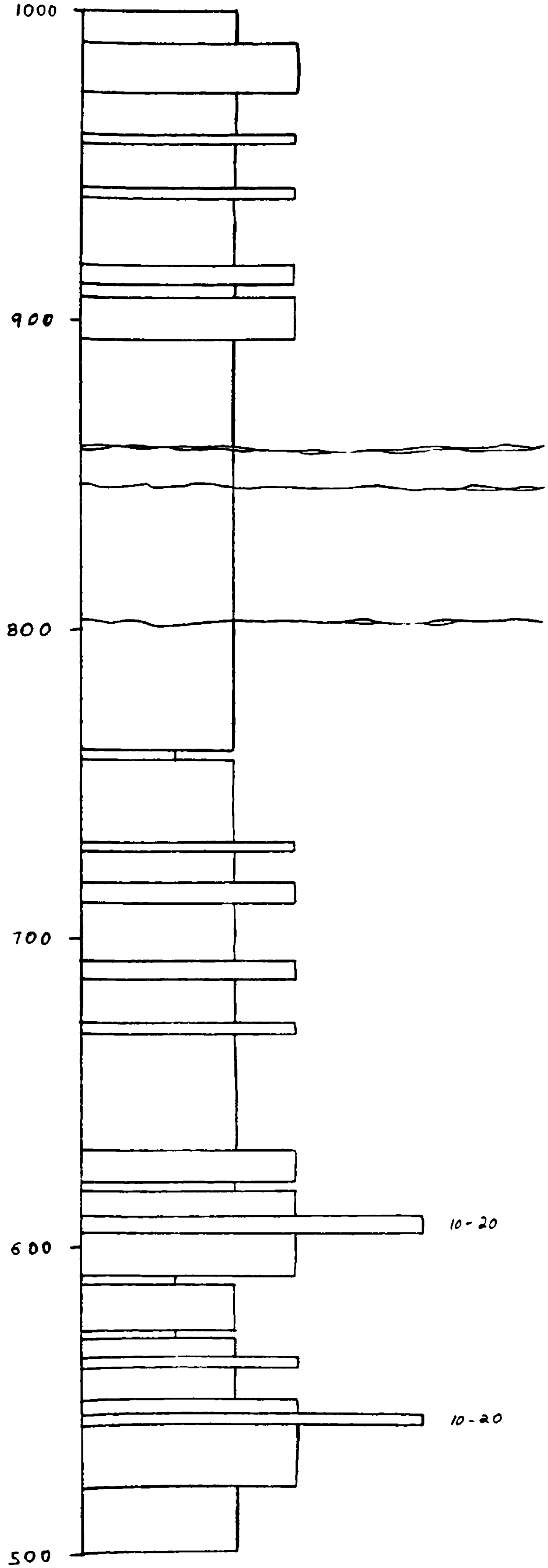

8000
7900
7800
St. Regis Fm.
7700
7600
7500


NCC
8500
8400
8300
St. Regis Fm.
8200
8100
Keybed :
Homogenous grey V.F. sand with large Load Structures,
Homogenous, green, clayey silty-sand.
Closely cleaved bed of 0.5 to 2 cm alpha couplets.
8000

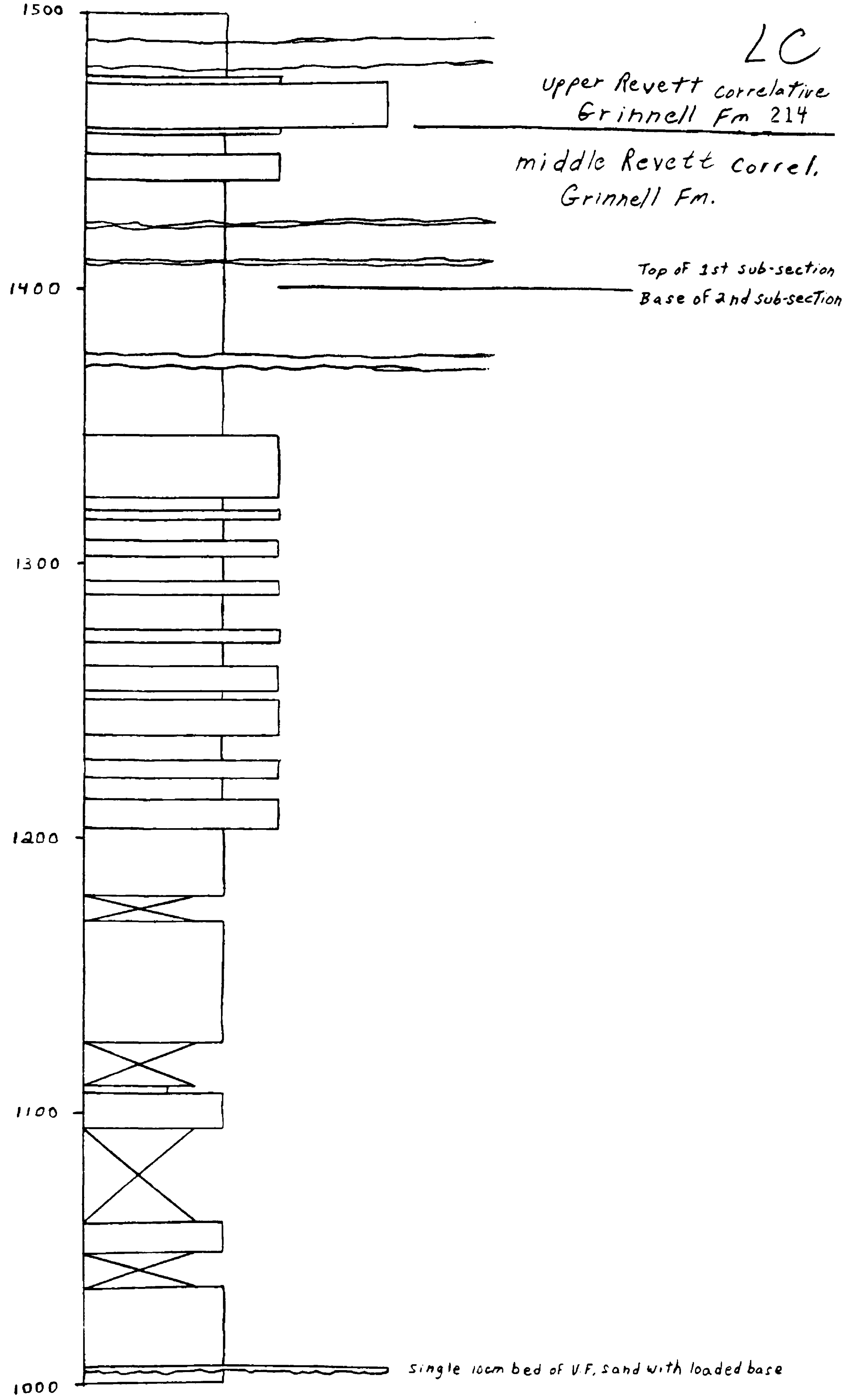

8700
Top of 5th sub-section
8600
Empire Fm.
St. Regis Fm.
8500


Lion Creek 212
500
400
300
200
100
0
middle Revett correl.
Grinnell Fm.
lower Revett correl.
Grinnell Fm.
20-30
20-40
20-30
20-60
10-30
30-40
15-30
30-50
20-30
10-15
10-15
10-15
10-15
10-20
Base of 1st
sub-section

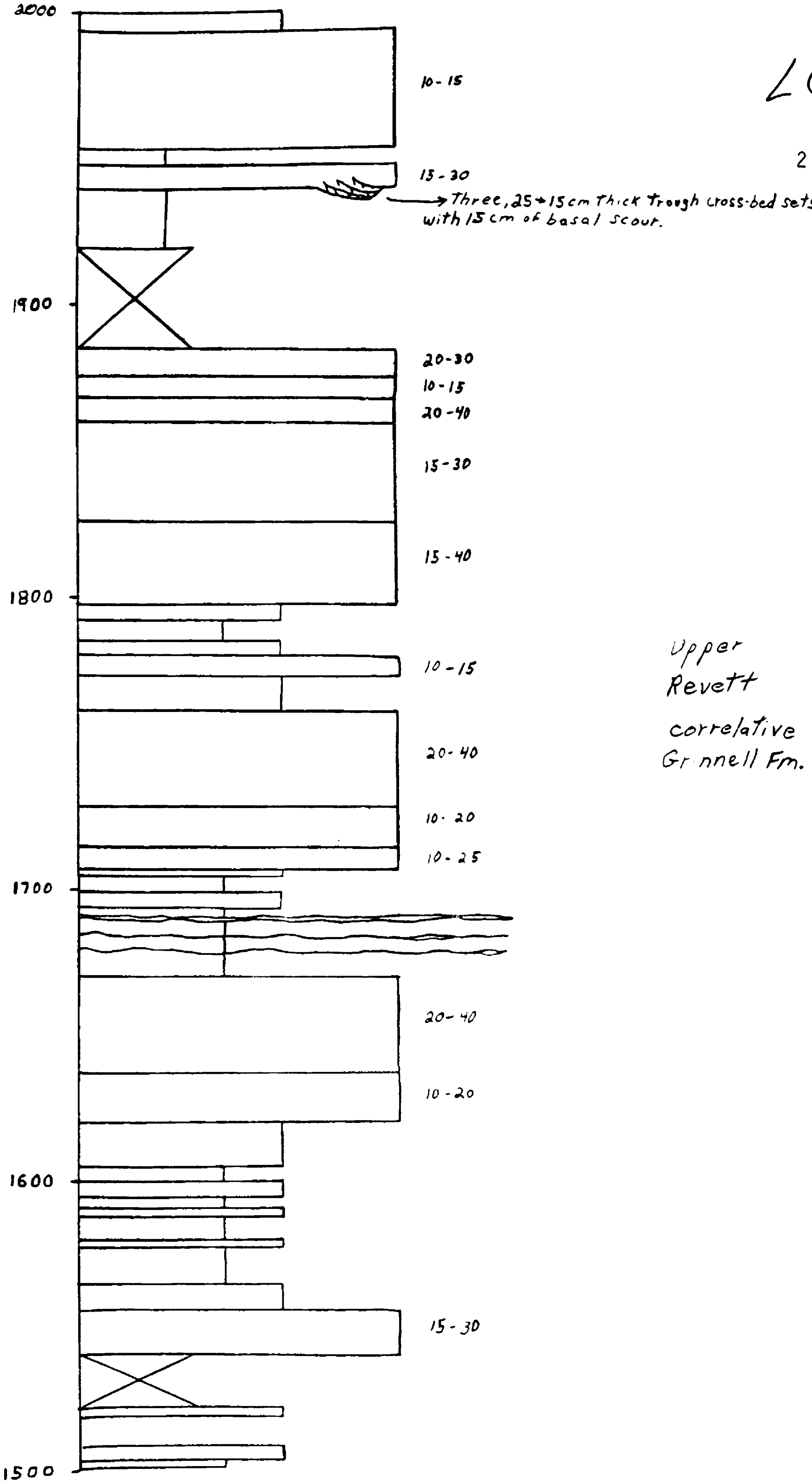
LC



middle Revett correlative Grinnell Fm.

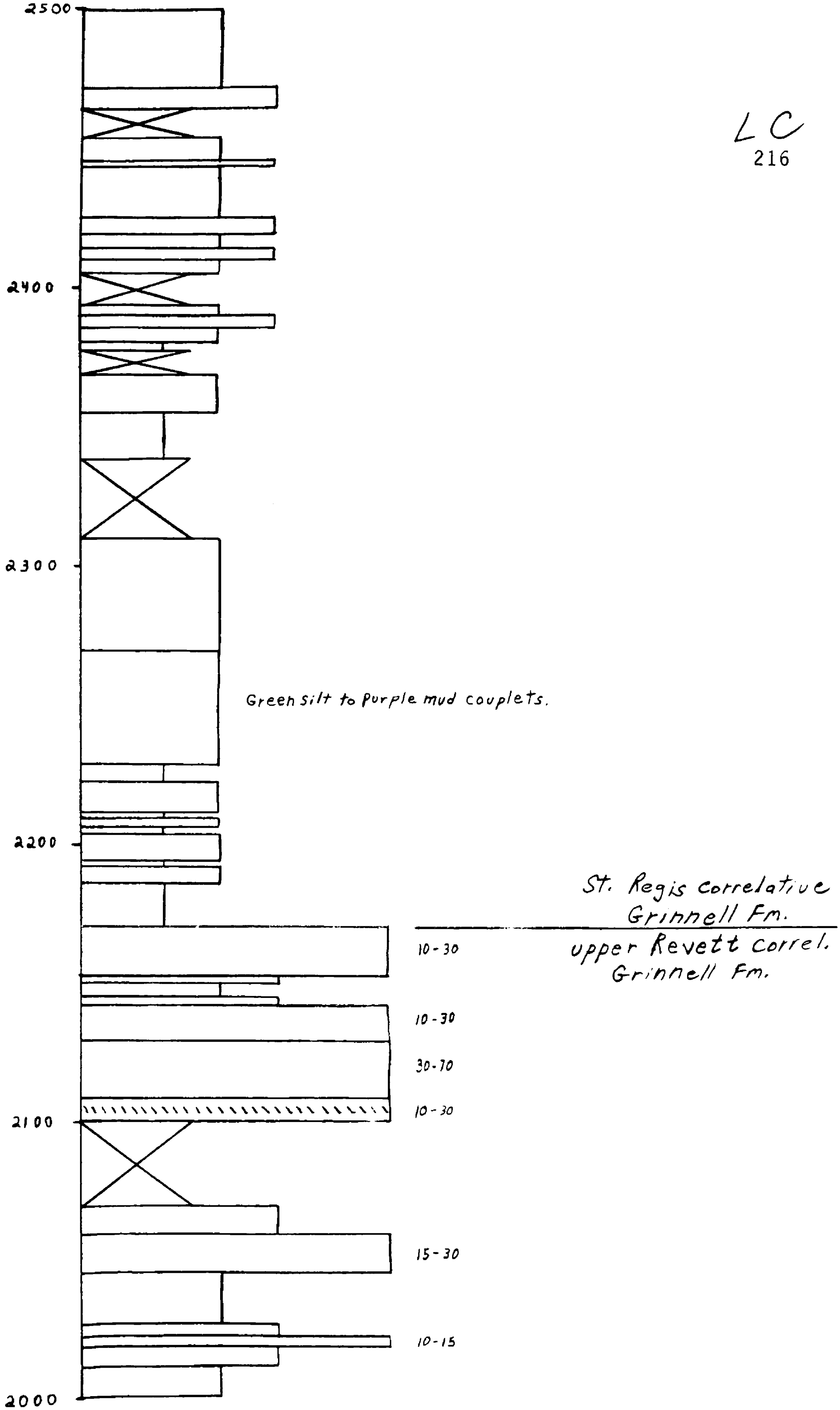

LC
Upper Revett correlative
Grinnell Fm 214
middle Revett correl.
Grinnell Fm.
Top of 1st sub-section
Base of 2nd sub-section
single 10cm bed of V.F. sand with loaded base
1500
1400
1300
1200
1100
1000

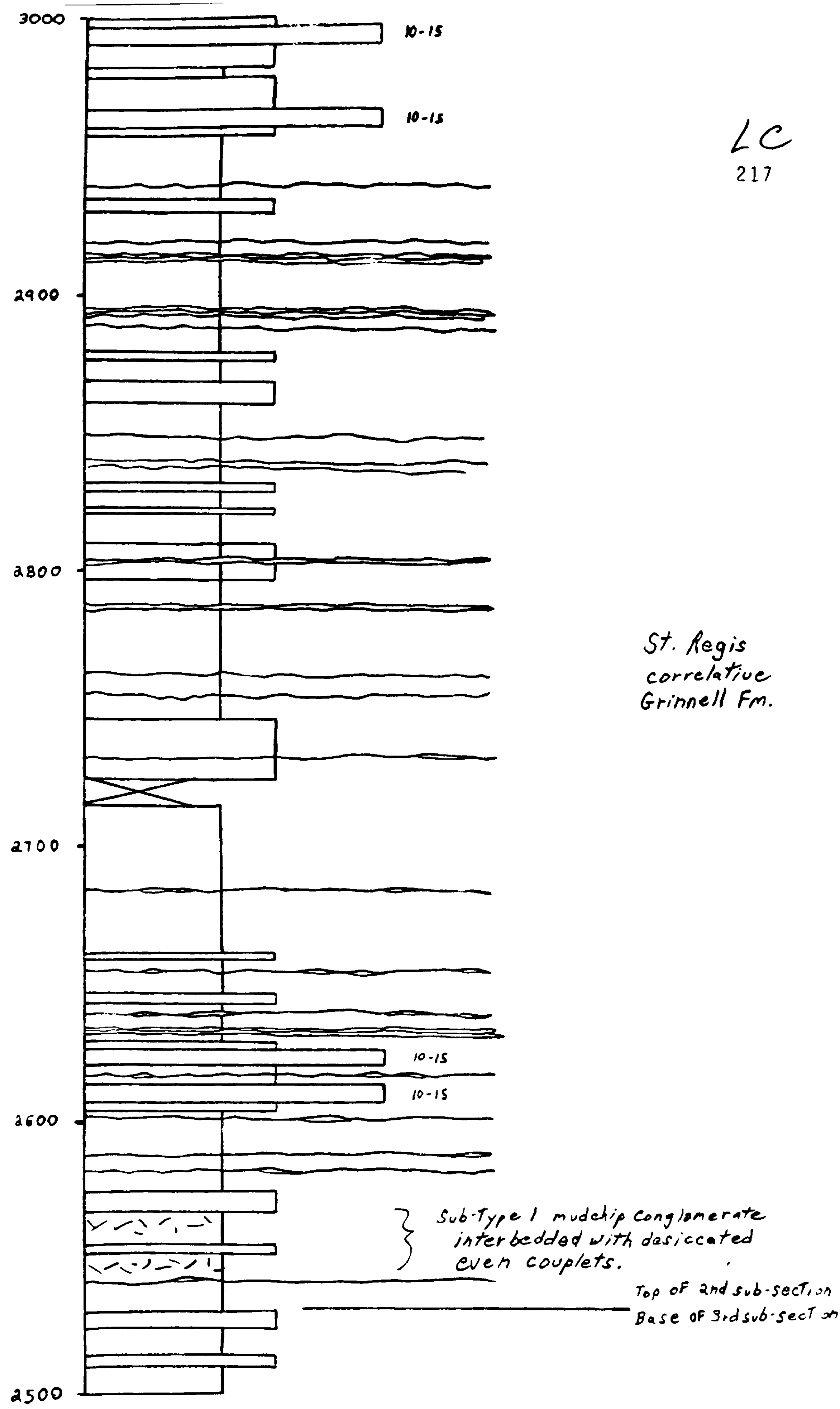
LC



Upper Revett correlative Grinnell Fm.

LC
216

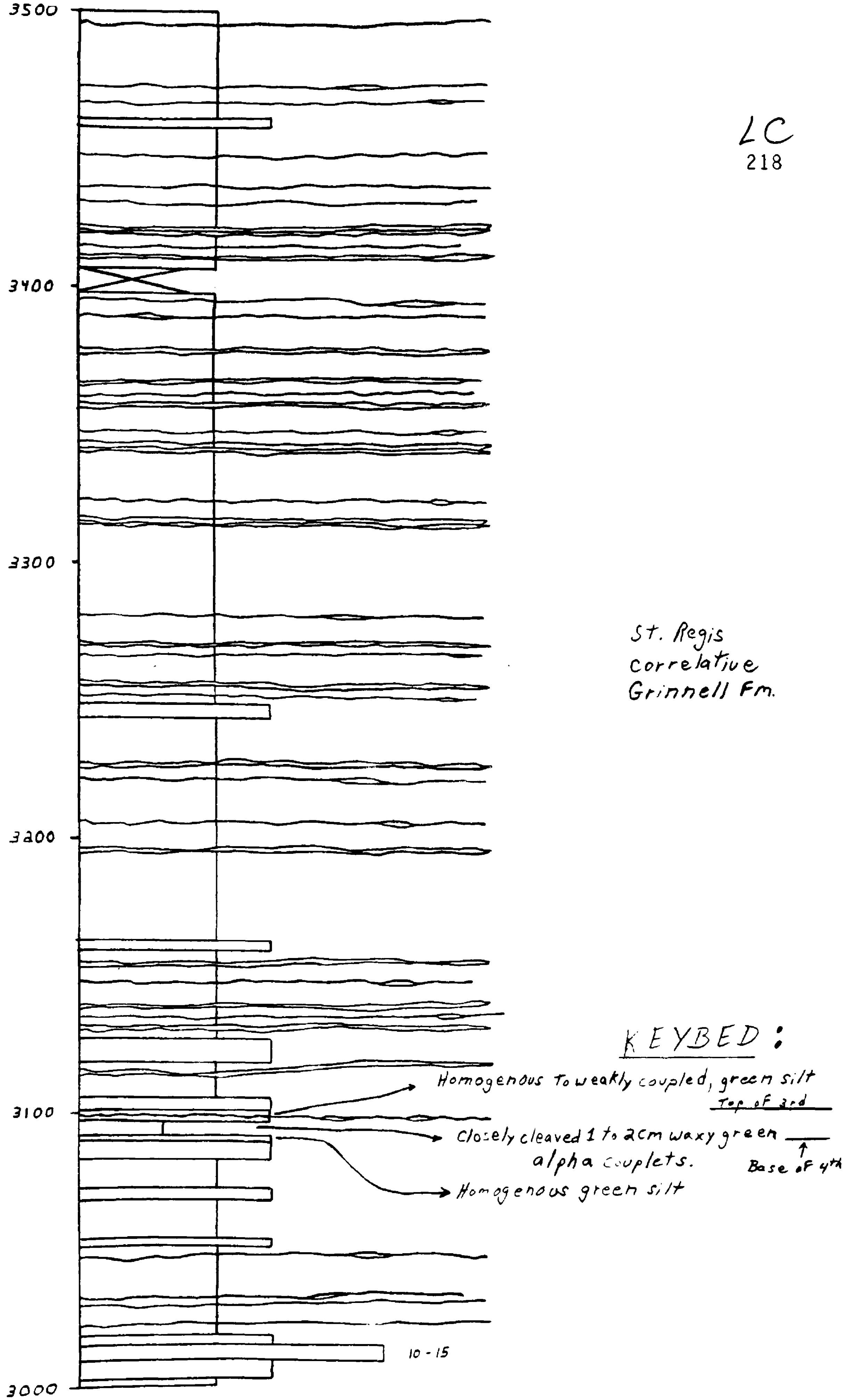

2500
2400
2300
2200
2100
2000
Green silt to purple mud couplets.
St. Regis correlative
Grinnell Fm.
upper Revett correl.
Grinnell Fm.
10-30
10-30
30-70
10-30
15-30
10-15

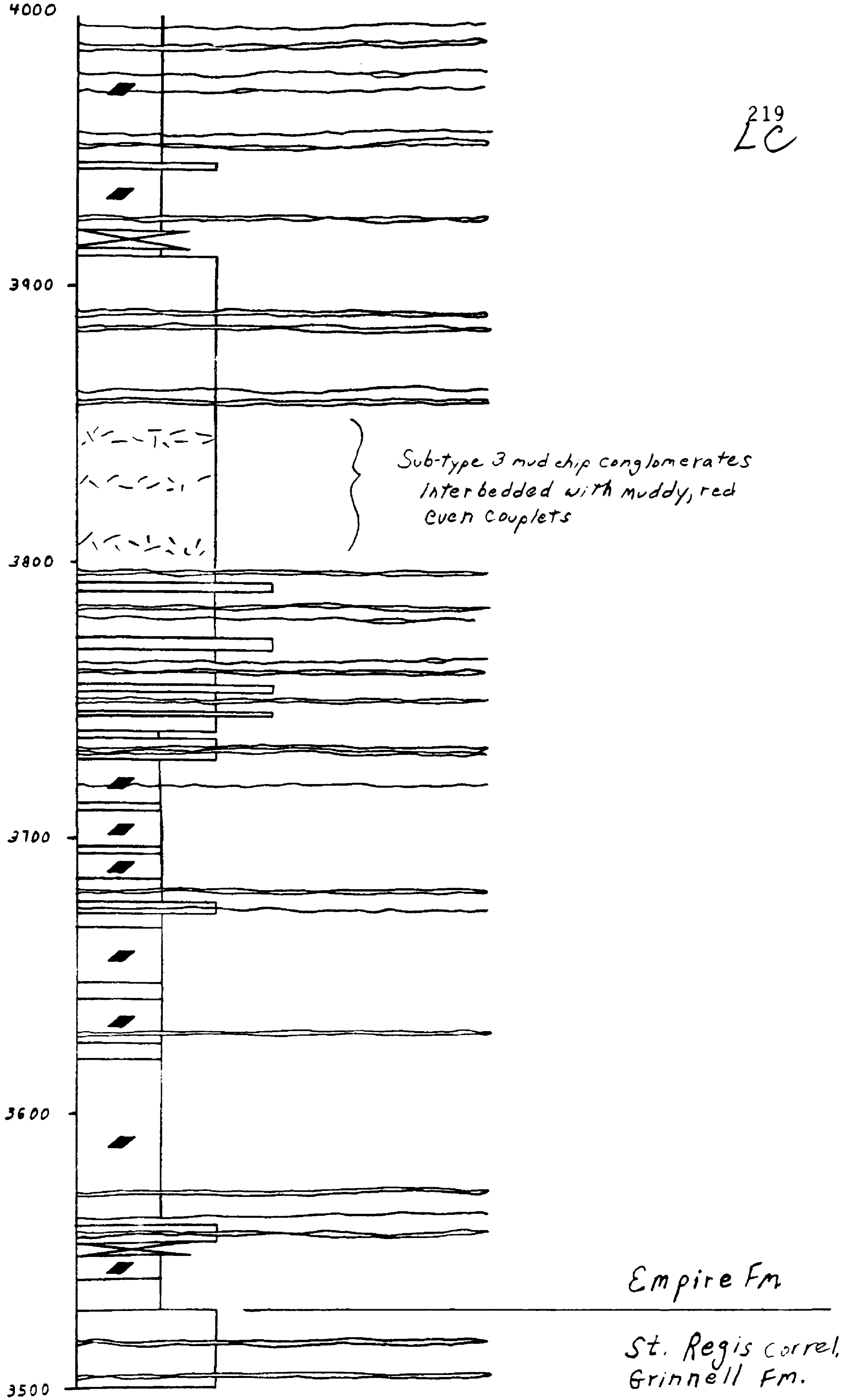


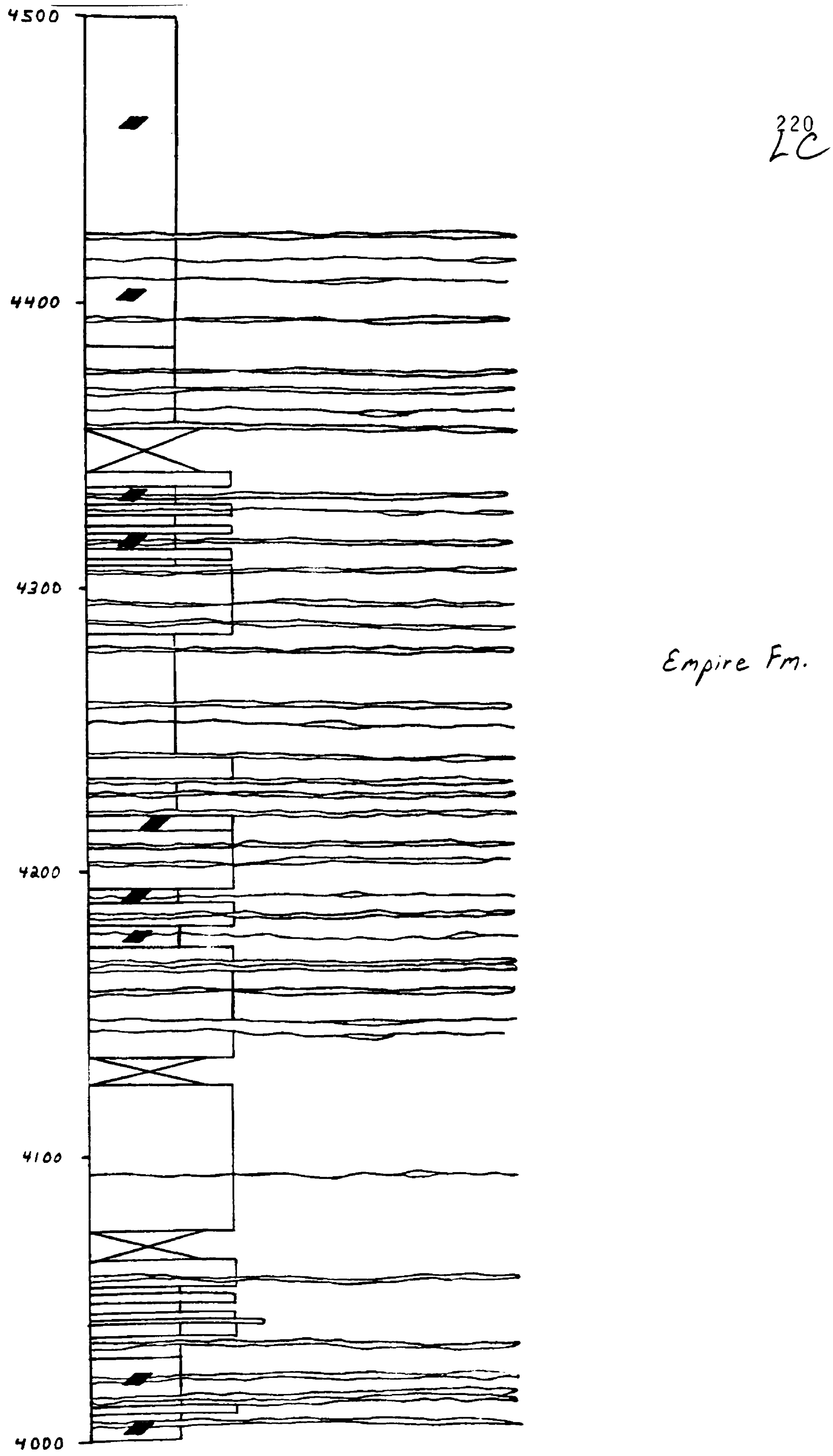
3000
10-15
10-15
2900
2800
2700
2600
10-15
10-15
2500
St. Regis
correlative
Grinnell Fm.
Sub-Type 1 mudchip conglomerate
interbedded with desiccated
even couplets.
Top of 2nd sub-section
Base of 3rd sub-section

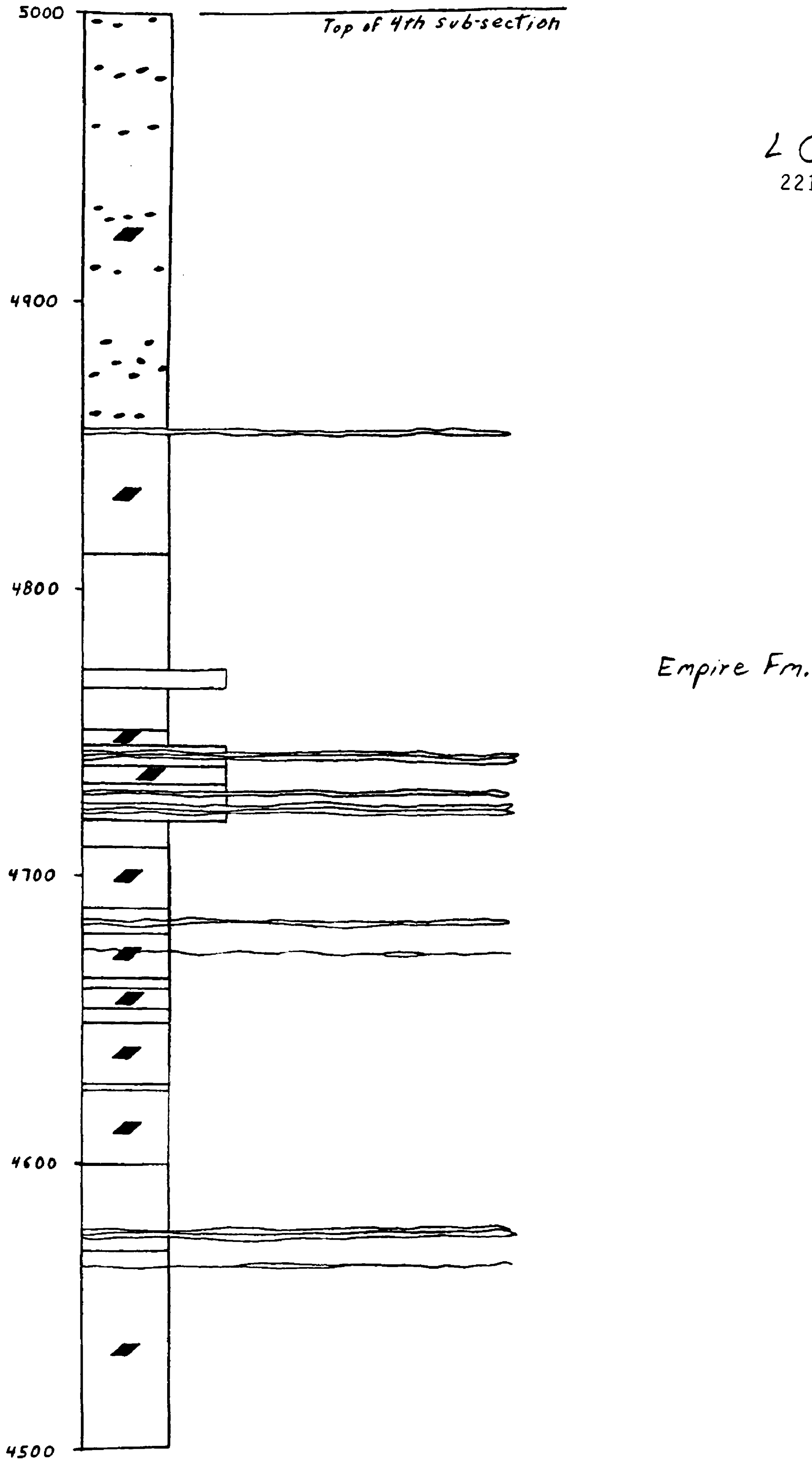
LC
217

LC
218

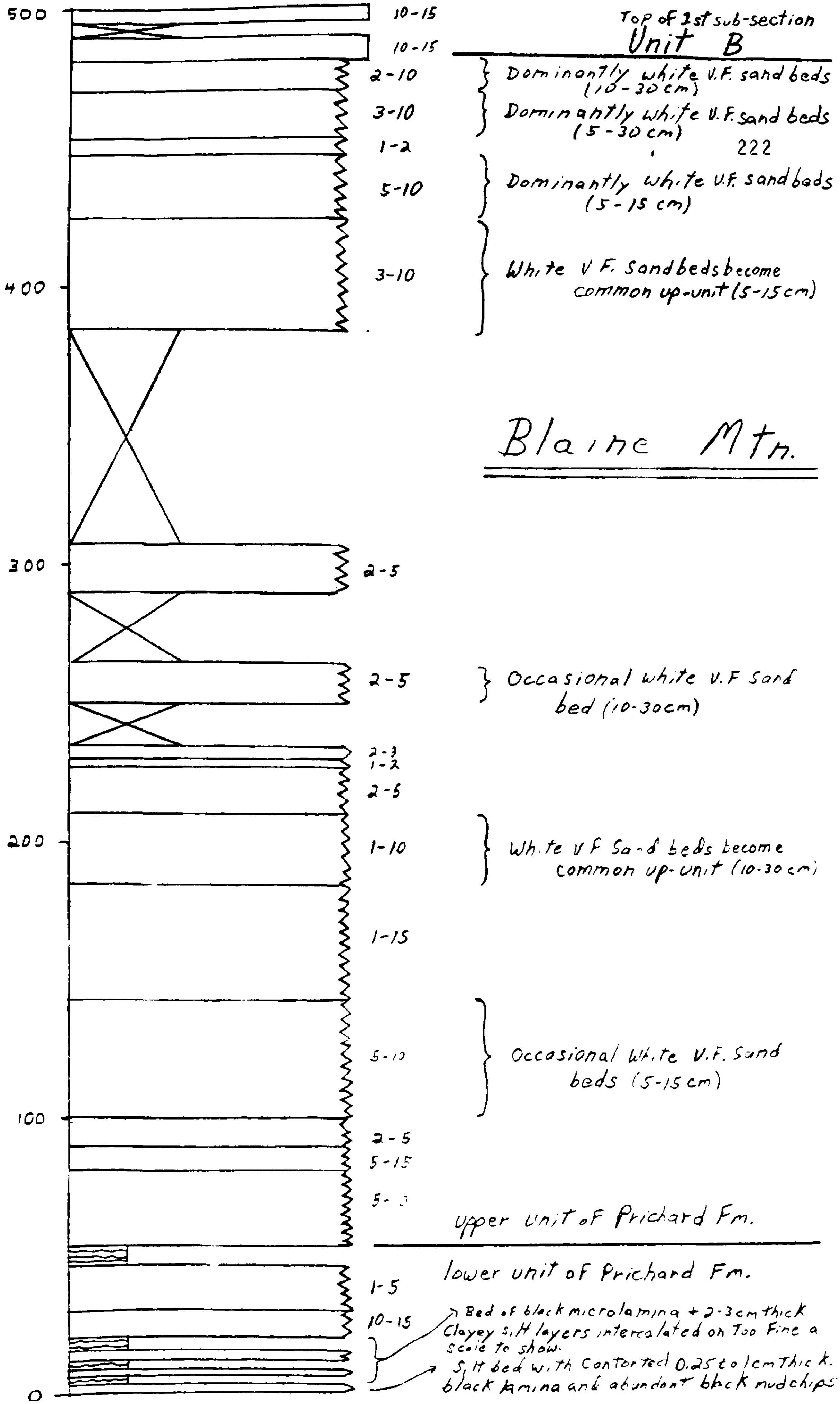

3500
3400
3300
3200
3100
3000
10 - 15
St. Regis
correlative
Grinnell Fm.
KEYBED:
Homogenous to weakly coupled, green silt
Top of 3rd
Closely cleaved 1 to 2cm waxy green
alpha couplets.
Base of 4th
Homogenous green silt


219
LC
4000
3900
Sub-type 3 mud chip conglomerates interbedded with muddy, red even couplets
3800
3700
3600
Empire Fm
St. Regis correl. Grinnell Fm.
3500

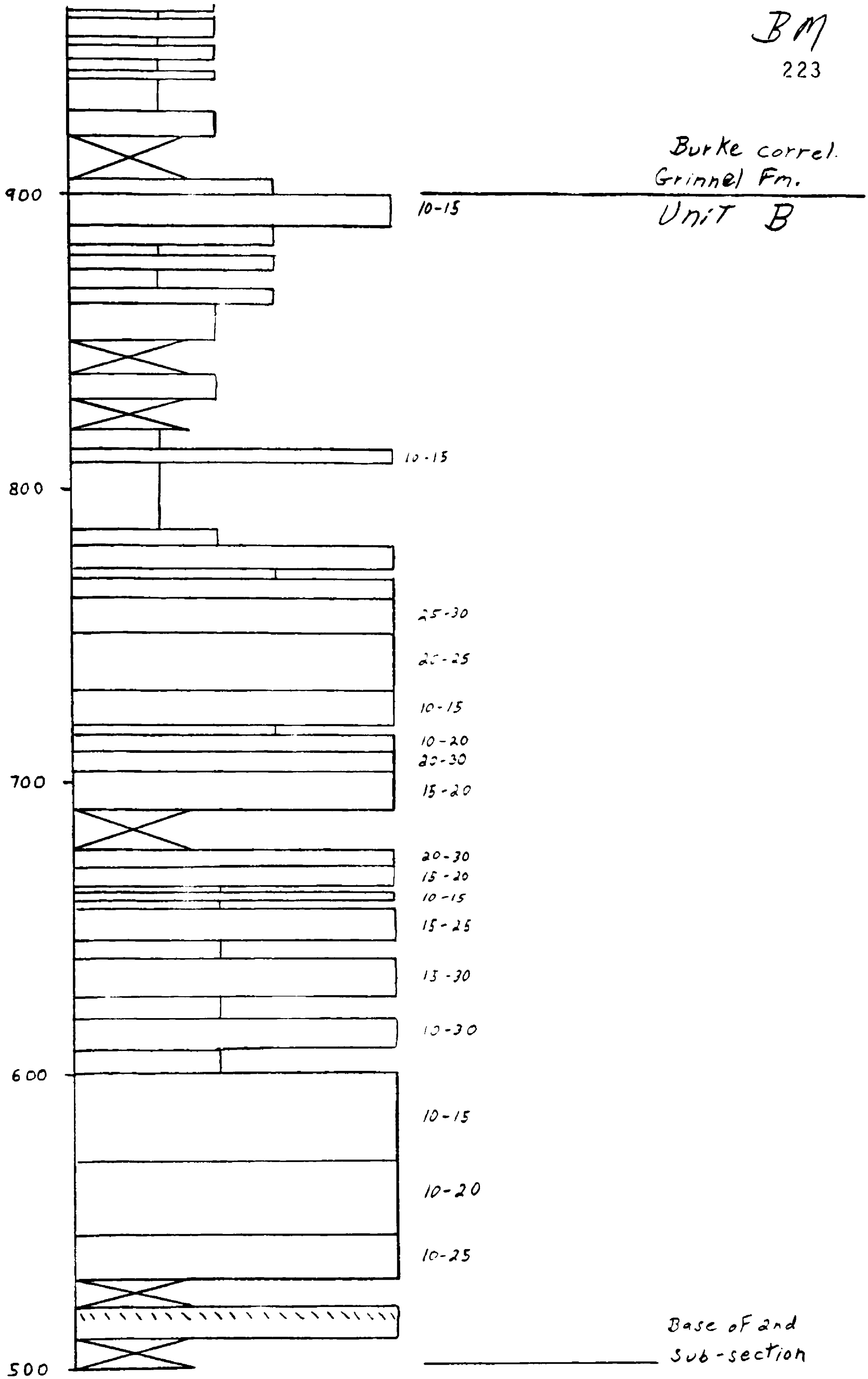

220
LC
4500
4400
4300
4200
4100
4000
Empire Fm.


5000
Top of 4th sub-section
4900
4800
4700
4600
4500
Empire Fm.

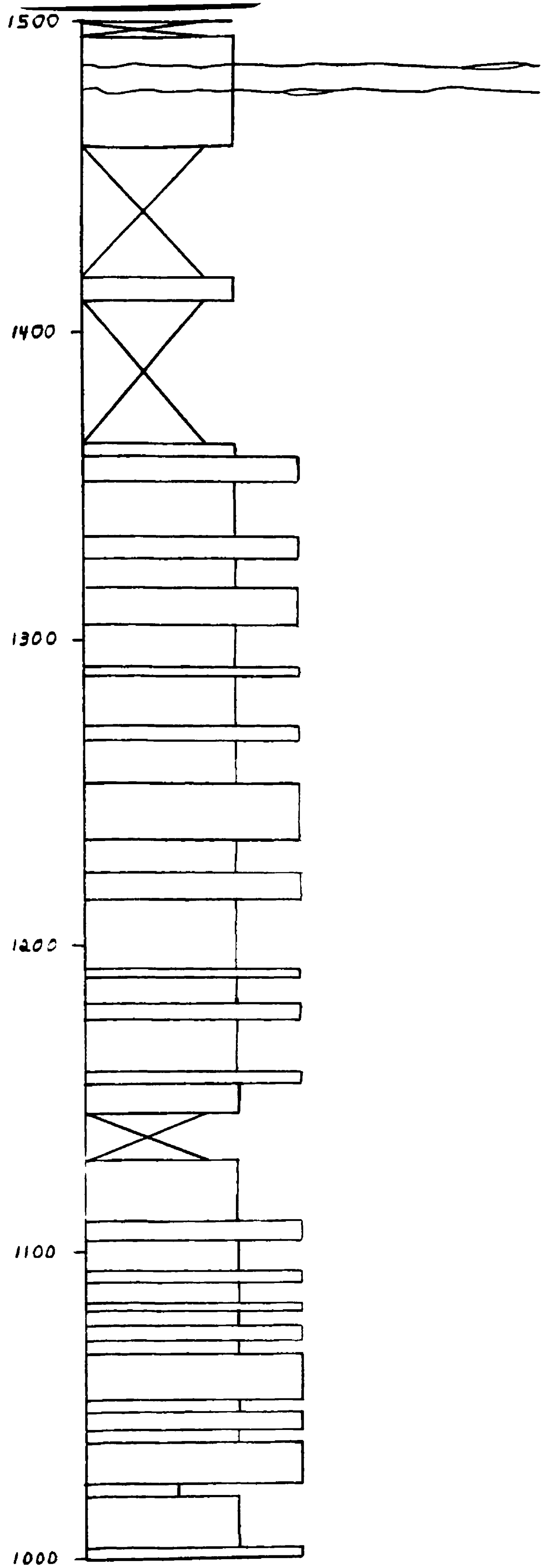
LC
221


500
400
300
200
100
0
10-15
10-15
2-10
3-10
1-2
5-10
3-10
2-5
2-5
2-3
1-2
2-5
1-10
1-15
5-10
2-5
5-15
1-5
10-15
Top of 1st sub-section
Unit B
Dominantly white V.F. sand beds (10-30 cm)
Dominantly white V.F. sand beds (5-30 cm)
Dominantly white V.F. sand beds (5-15 cm)
White V.F. Sandbeds become common up-unit (5-15 cm)
Blaine Mtn.
Occasional white V.F Sand bed (10-30cm)
White V.F Sand beds become common up-unit (10-30 cm)
Occasional white V.F. Sand beds (5-15 cm)
upper unit of Prichard Fm.
lower unit of Prichard Fm.
Bed of black microlamina + 2-3 cm thick Clayey silt layers intercalated on Too Fine a Scale to show.
Silt bed with Contorted 0.25 to 1cm Thick black lamina and abundant black mudchips.

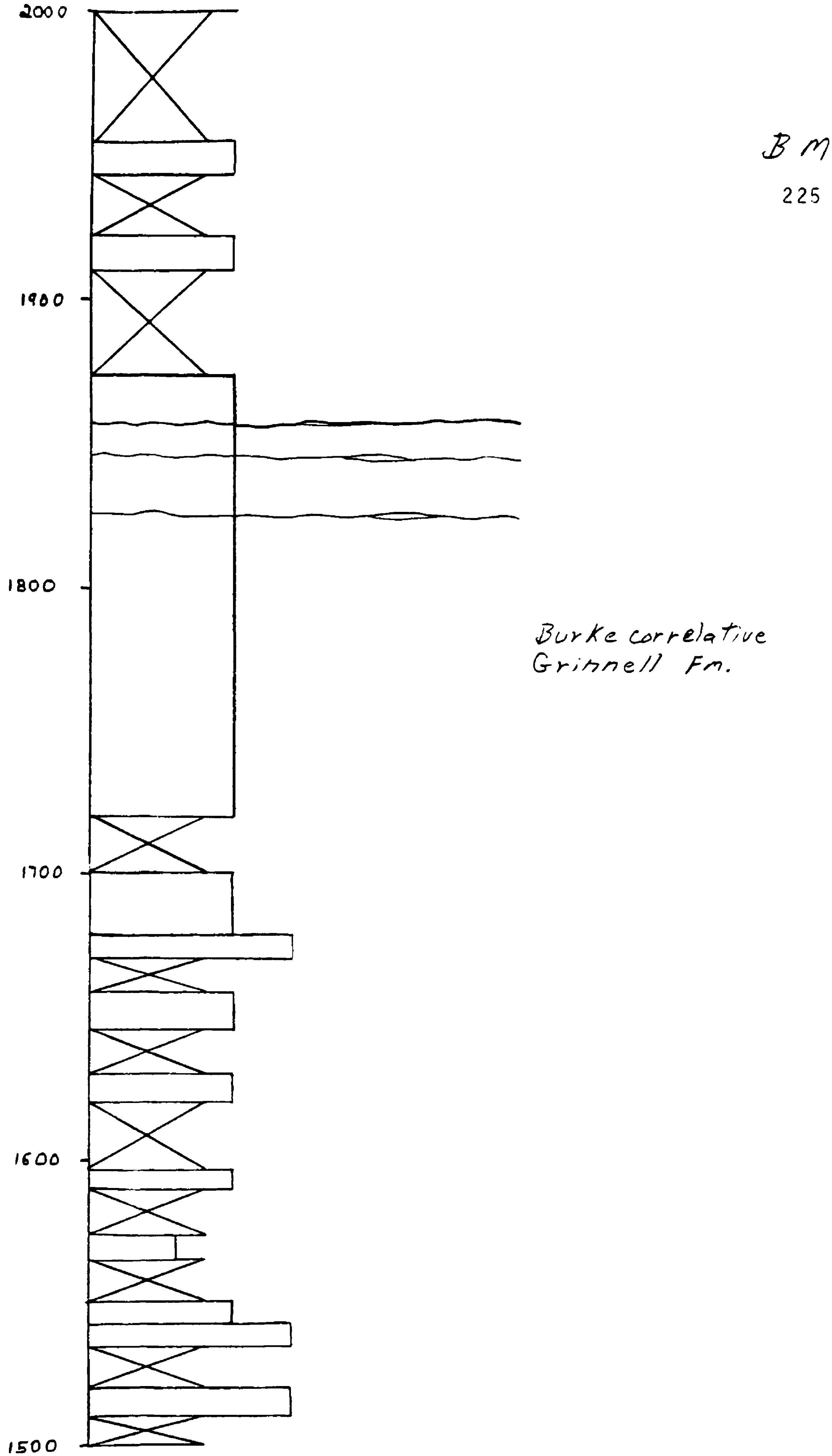
BM
223


Burke correl.
Grinnel Fm.
Unit B
900
10-15
10-15
800
25-30
20-25
10-15
10-20
20-30
700
15-20
20-30
15-20
10-15
15-25
13-30
10-30
600
10-15
10-20
10-25
Base of 2nd
Sub-section
500

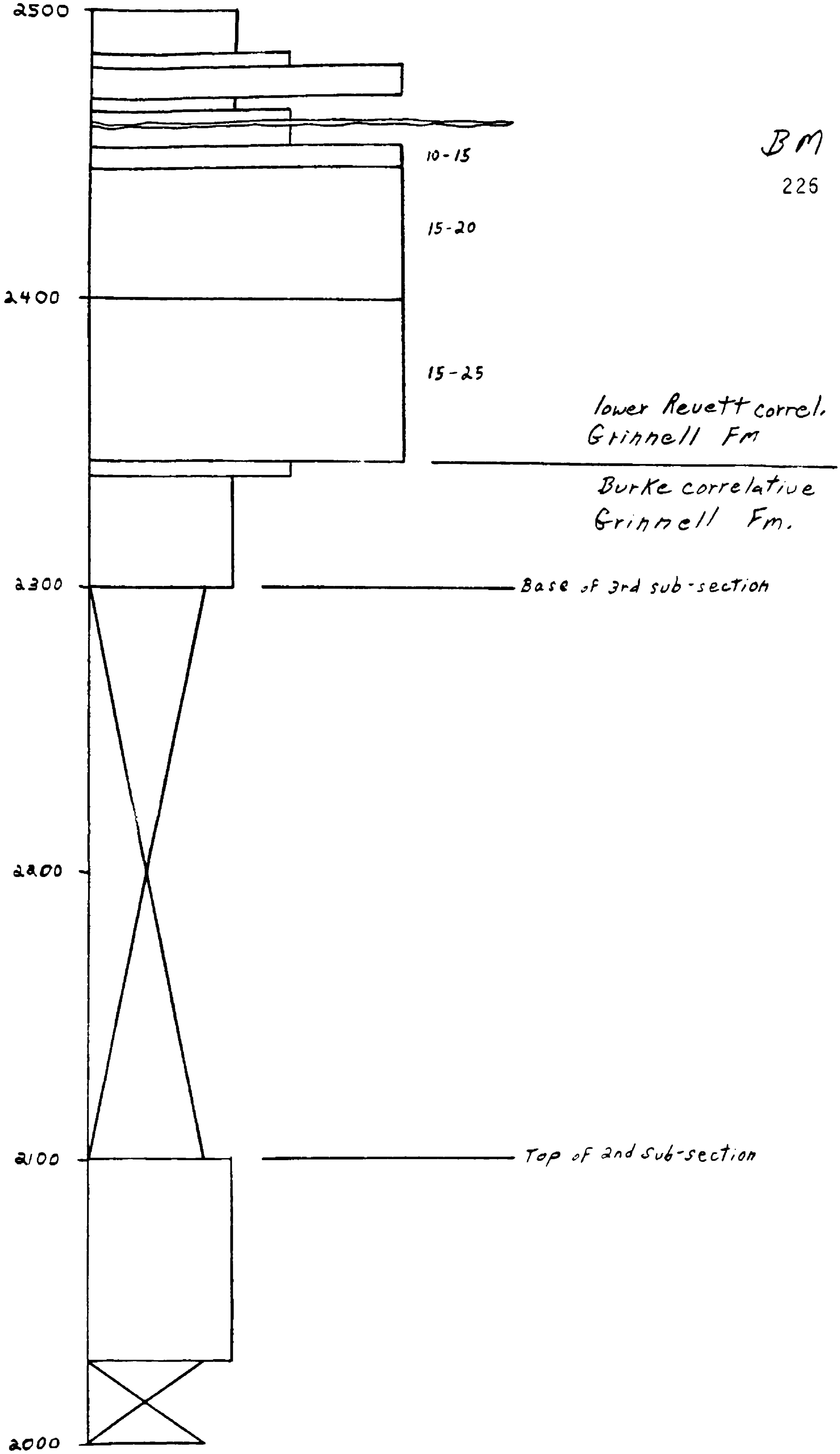
BM
224



Burke correlative
Grinnell Fm.

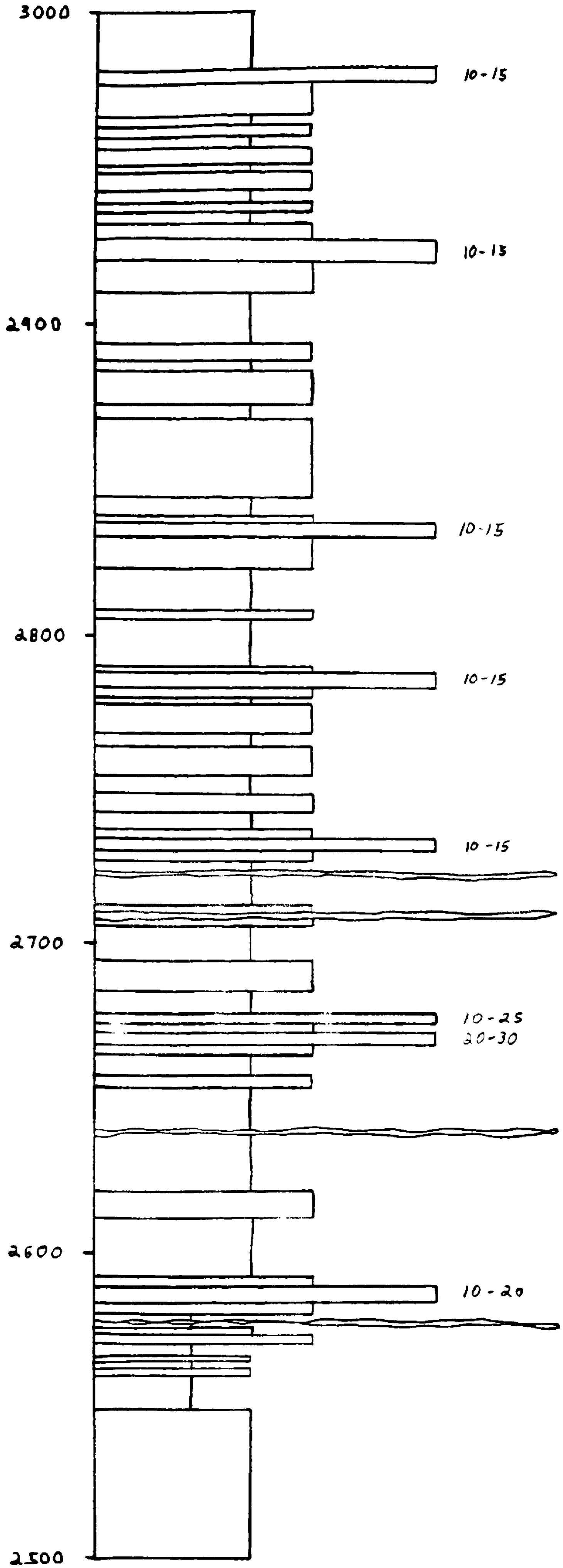

B M
225
2000
1900
1800
1700
1600
1500
Burke correlative
Grinnell Fm.

BM
226


2500
10-15
15-20
2400
15-25
lower Revett correl. Grinnell Fm
Burke correlative Grinnell Fm.
2300
Base of 3rd sub-section
2200
2100
Top of 2nd sub-section
2000

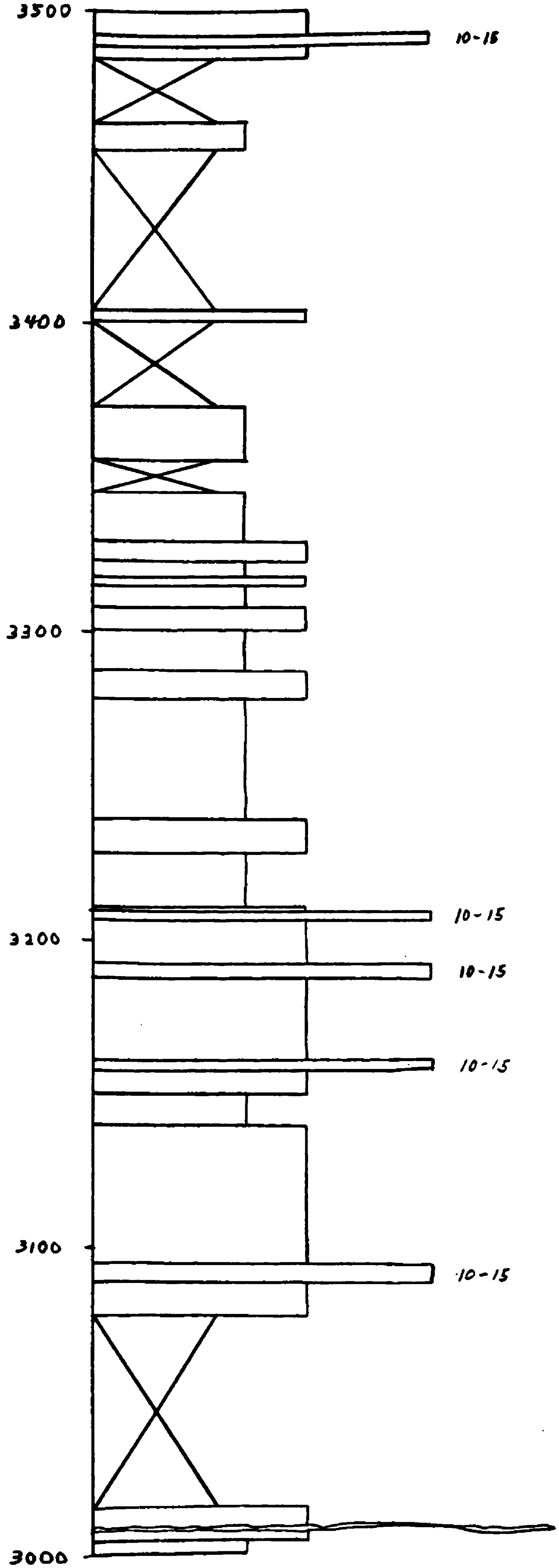
BM

227



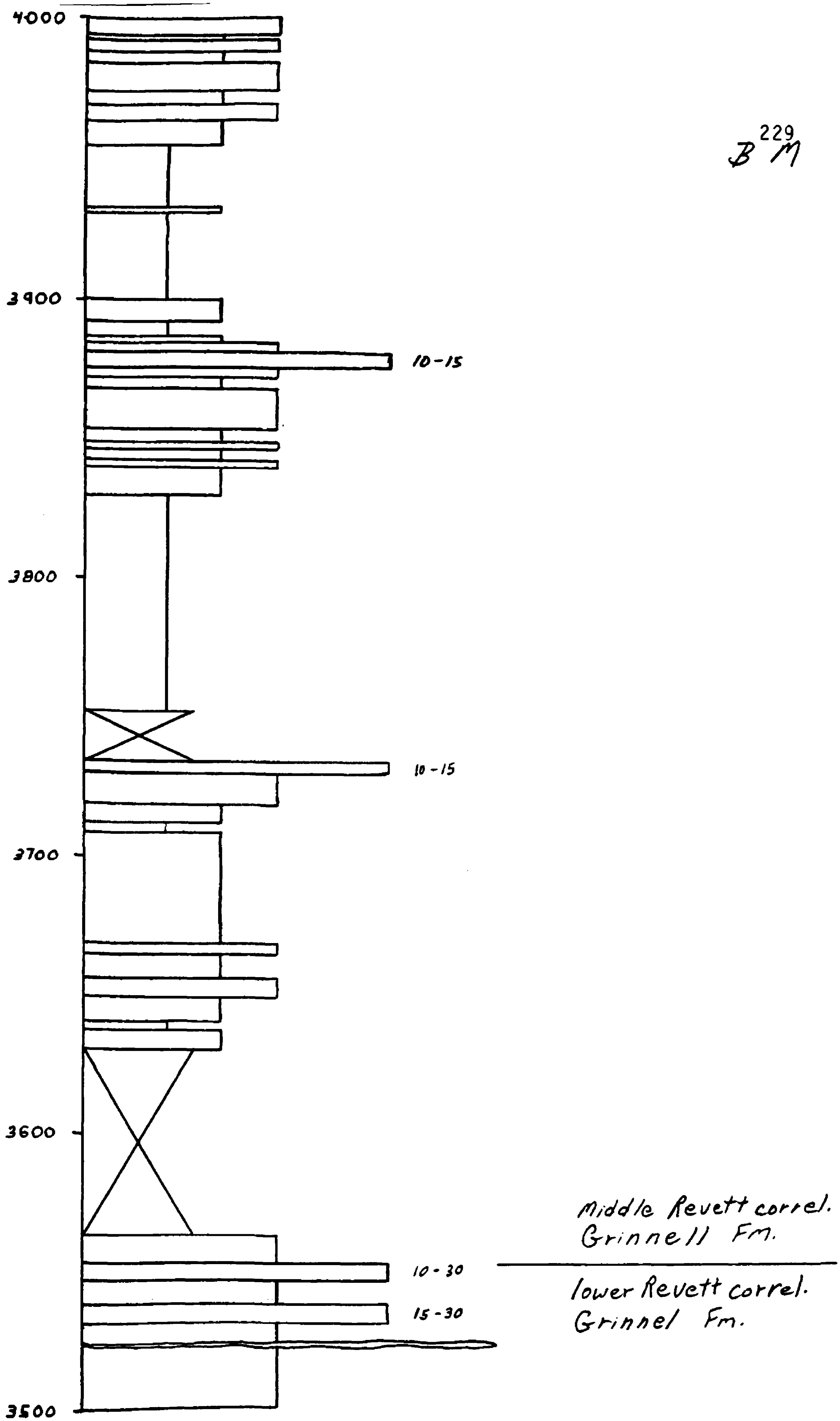
lower Revett correlative
Grinnell Fm.

BM



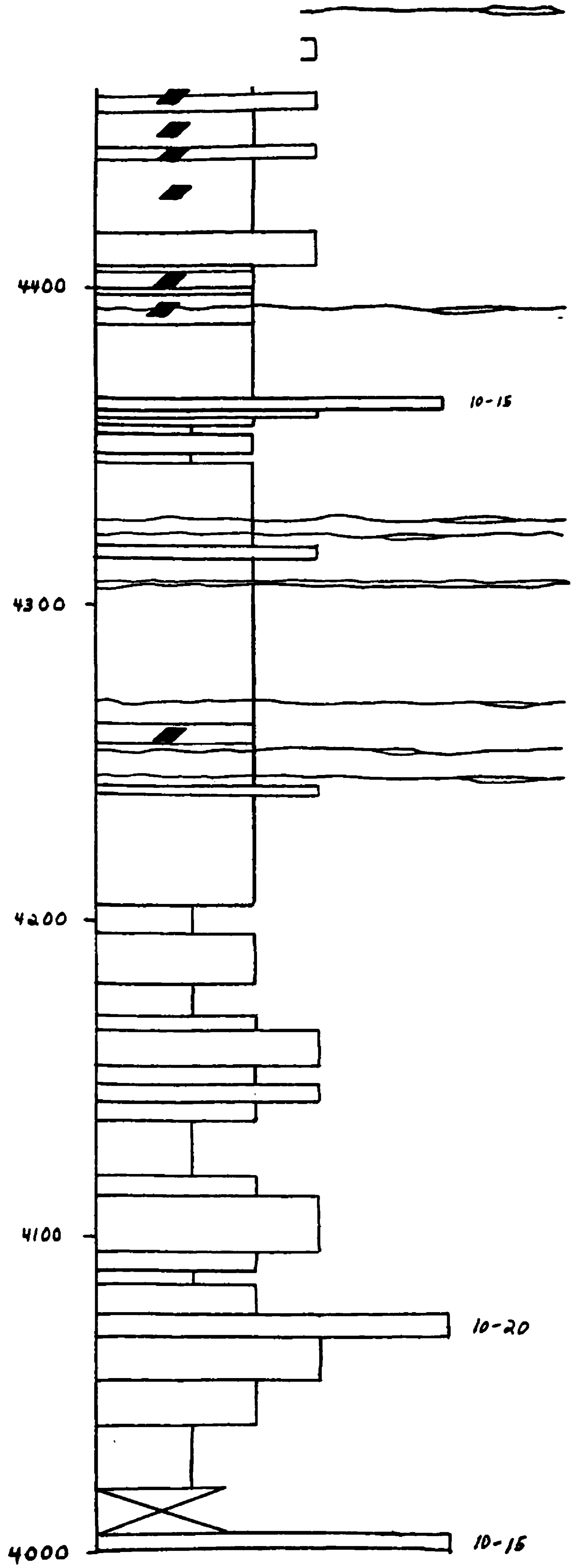
lower Revett Correlative
Grinnell Fm.

B 229 M



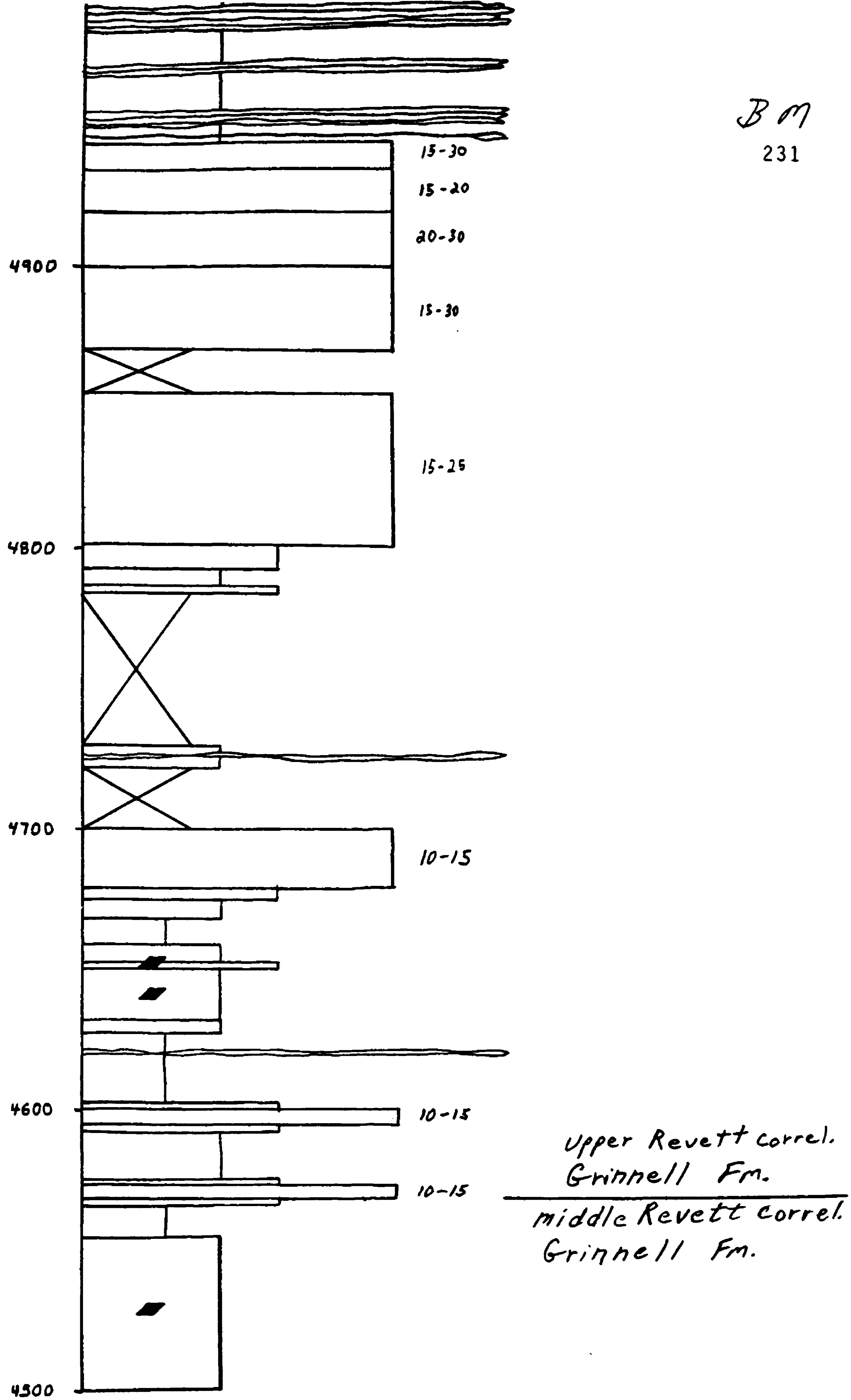
4000
3900
10-15
3800
10-15
3700
3600
Middle Revett correl.
Grinnell Fm.
10-30
lower Revett correl.
Grinnel Fm.
15-30
3500

BM
230

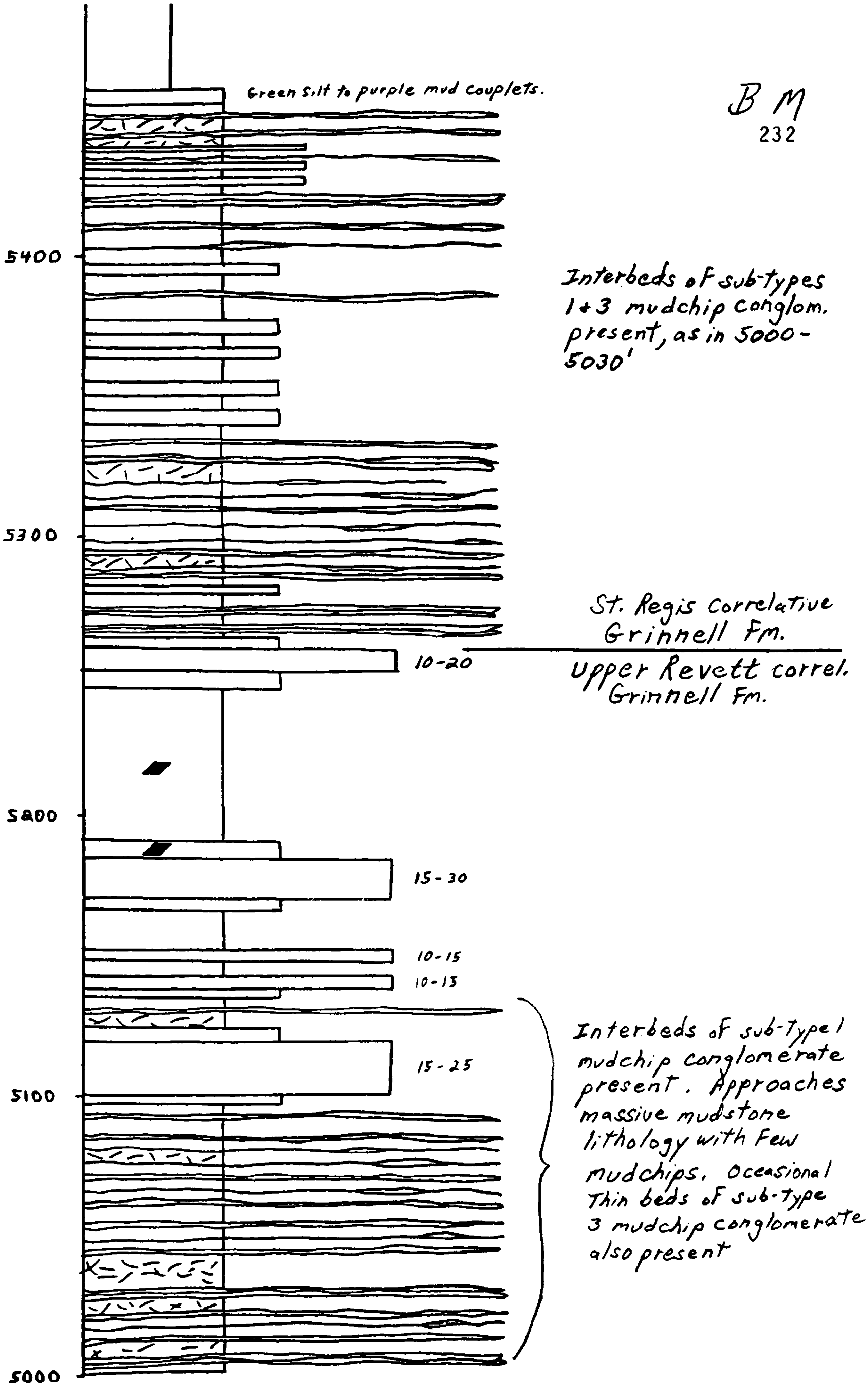


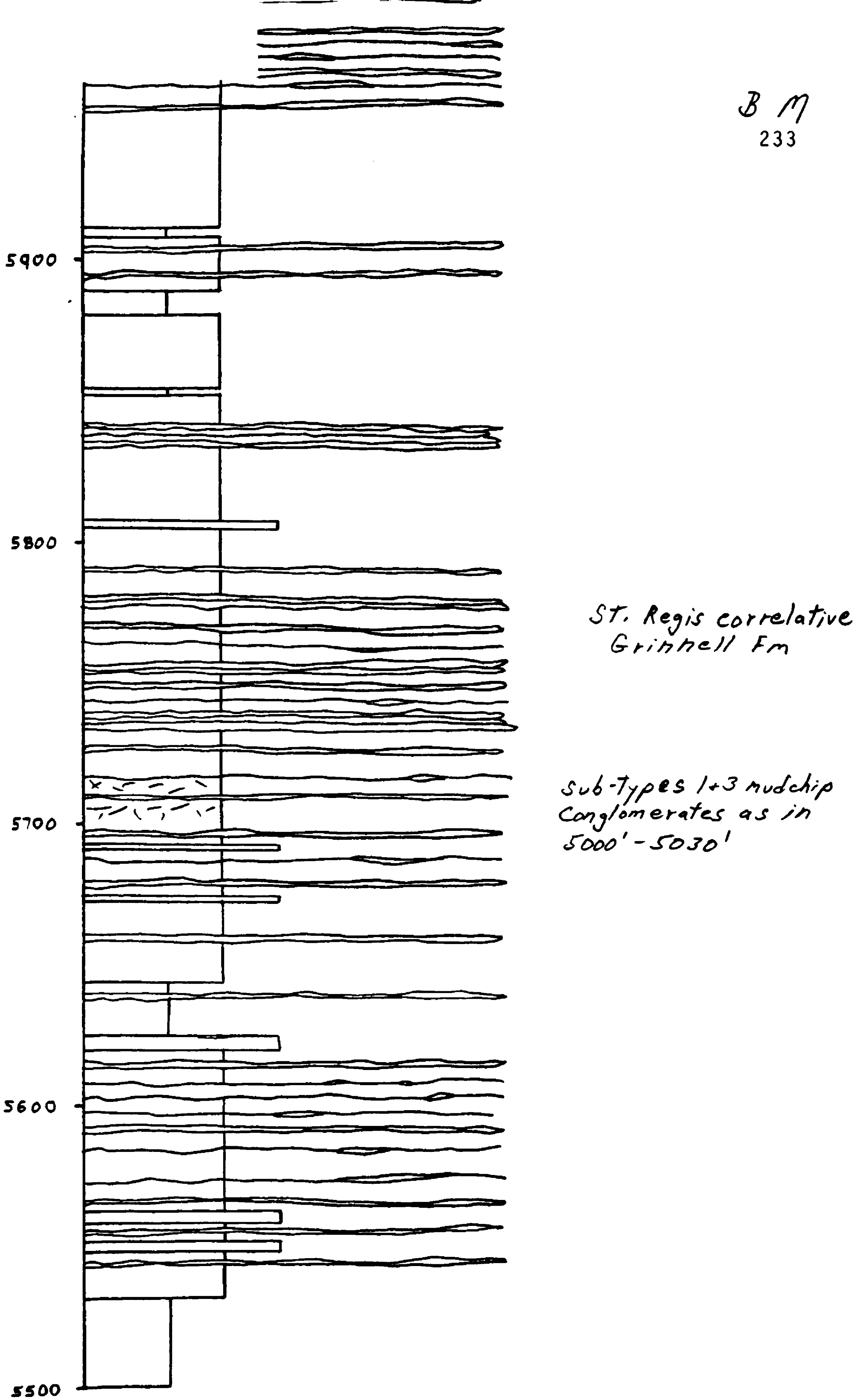
middle Revett correl.
Grinnell Fm.

BM
231

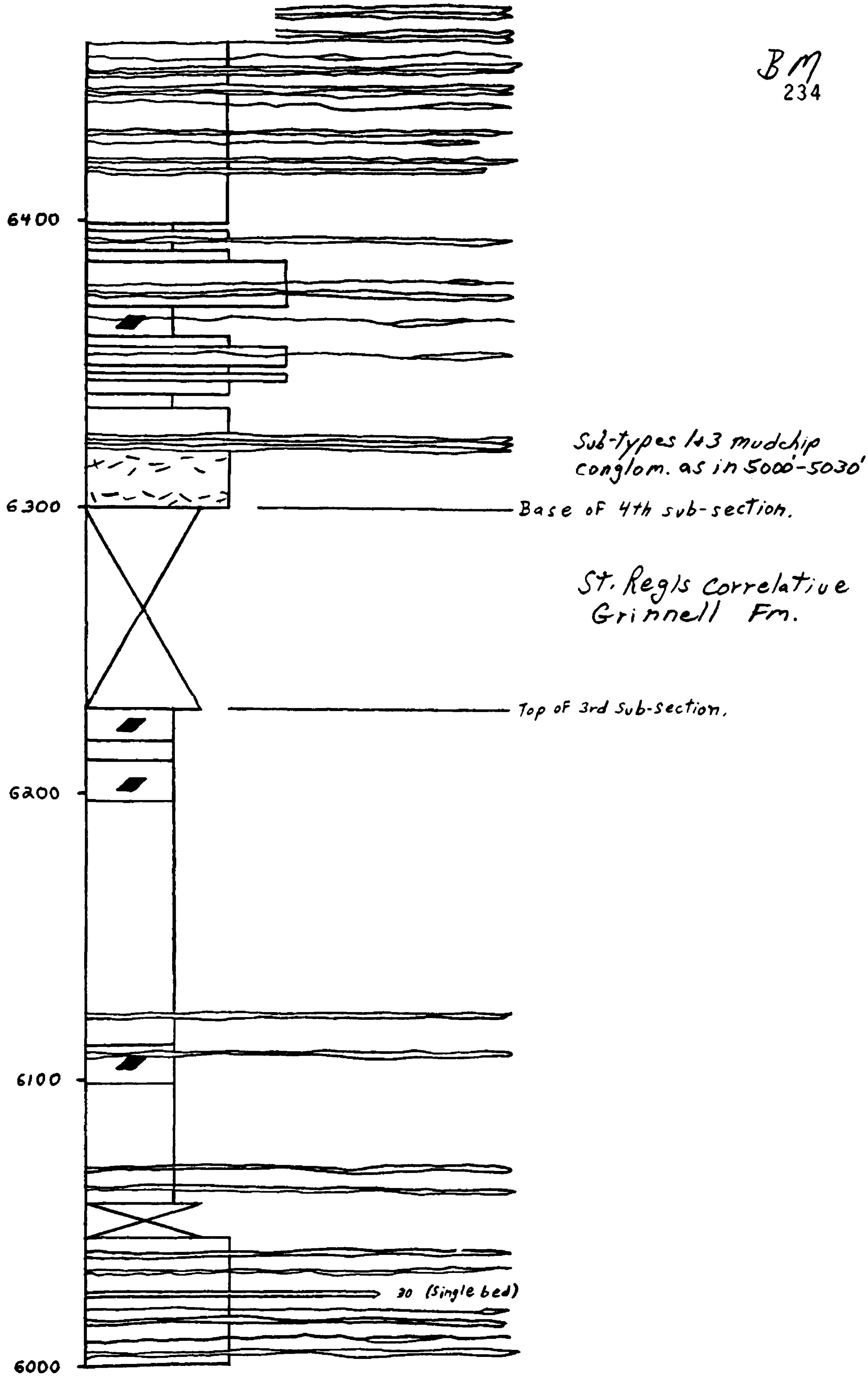


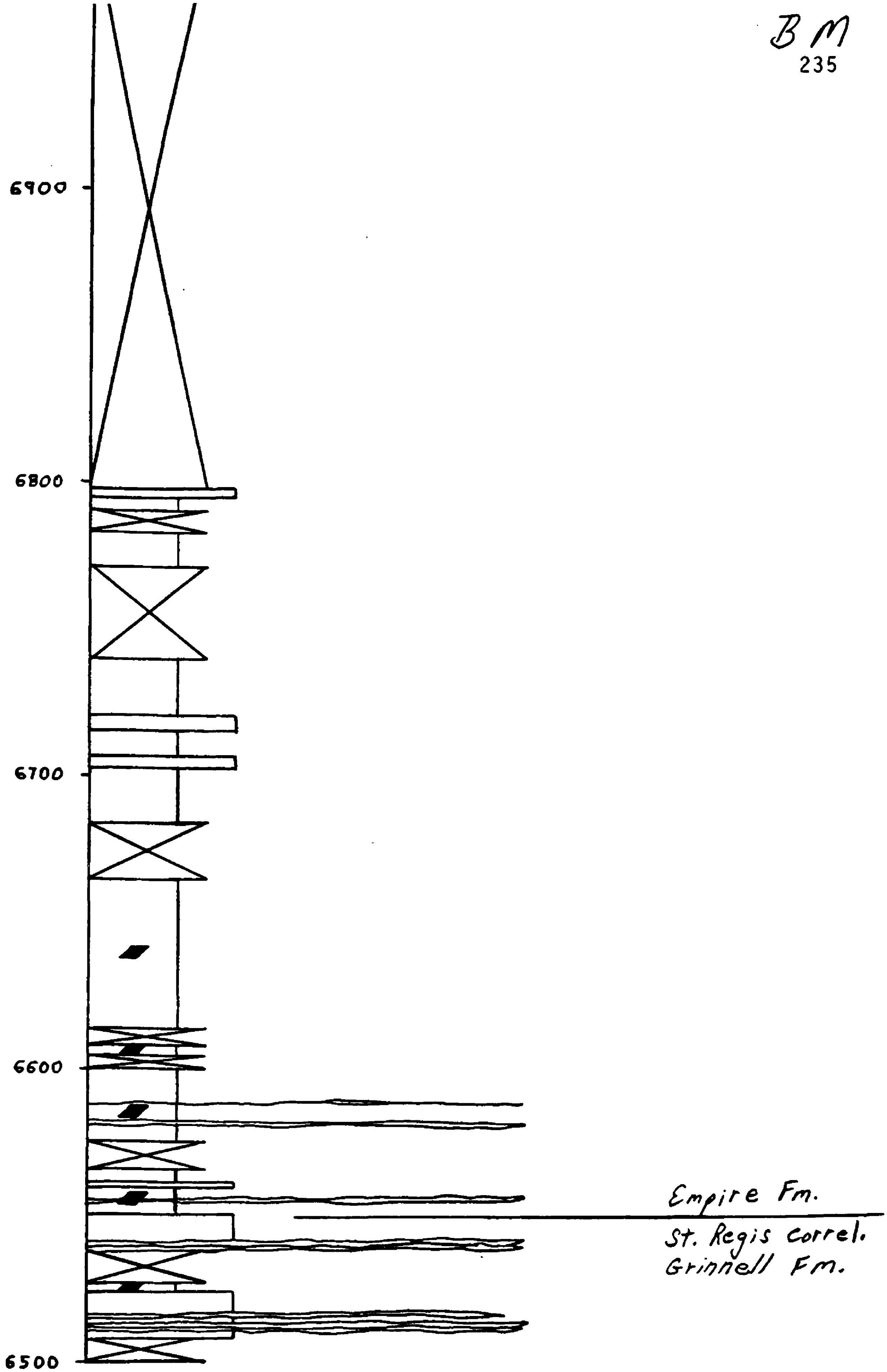
15-30
15-20
20-30
4900
15-30
15-25
4800
4700
10-15
4600
10-15
10-15
4500
Upper Revett correl.
Grinnell Fm.
middle Revett correl.
Grinnell Fm.

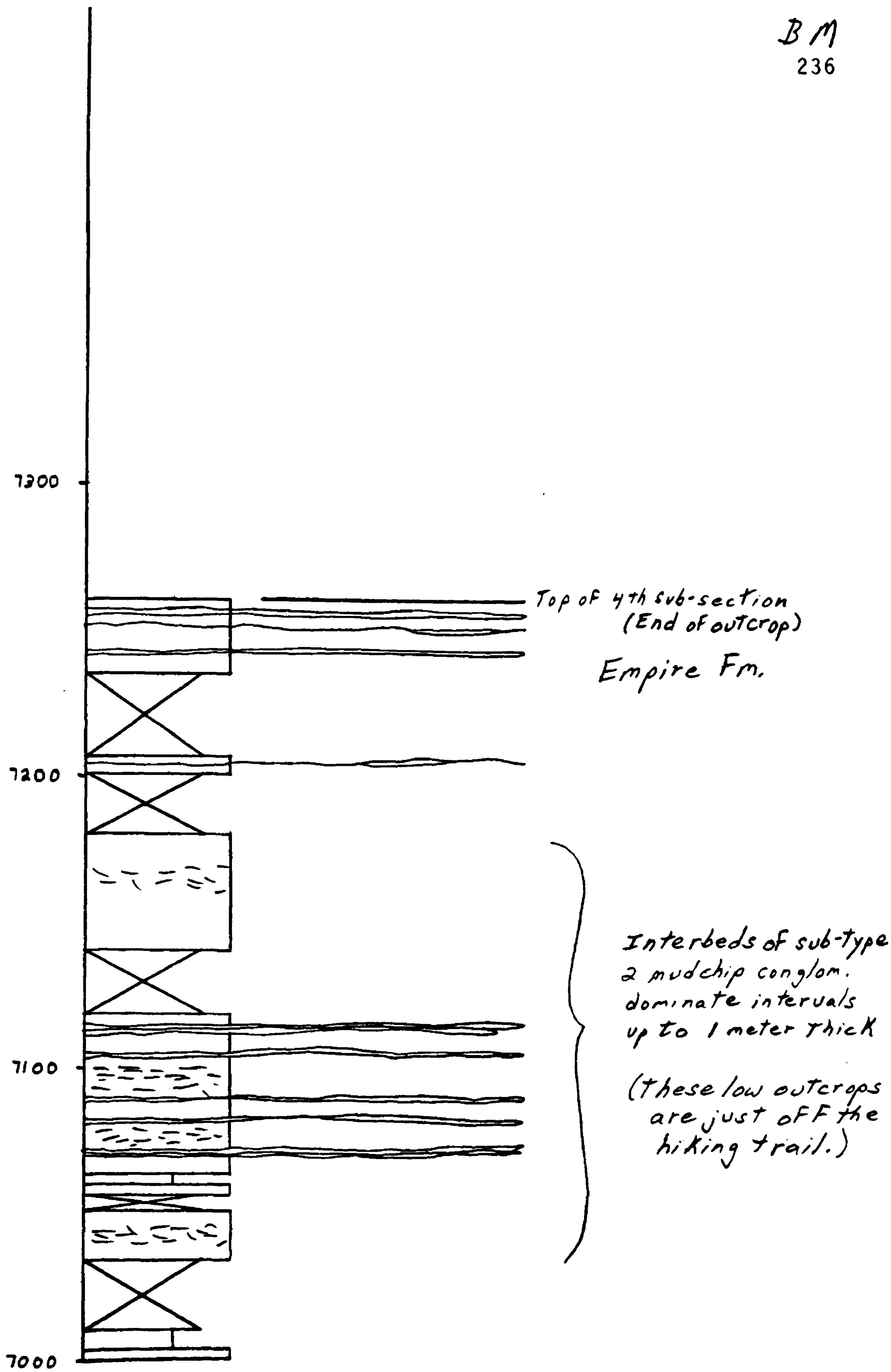

B M
232
Green silt to purple mud couplets.
5400
Interbeds of sub-types 1+3 mudchip conglom. present, as in 5000-5030'
5300
St. Regis Correlative Grinnell Fm.
10-20
Upper Revett correl. Grinnell Fm.
5200
15-30
10-15
10-13
15-25
Interbeds of sub-Type 1 mudchip conglomerate present. Approaches massive mudstone lithology with few mudchips. Occasional thin beds of sub-type 3 mudchip conglomerate also present
5100
5000



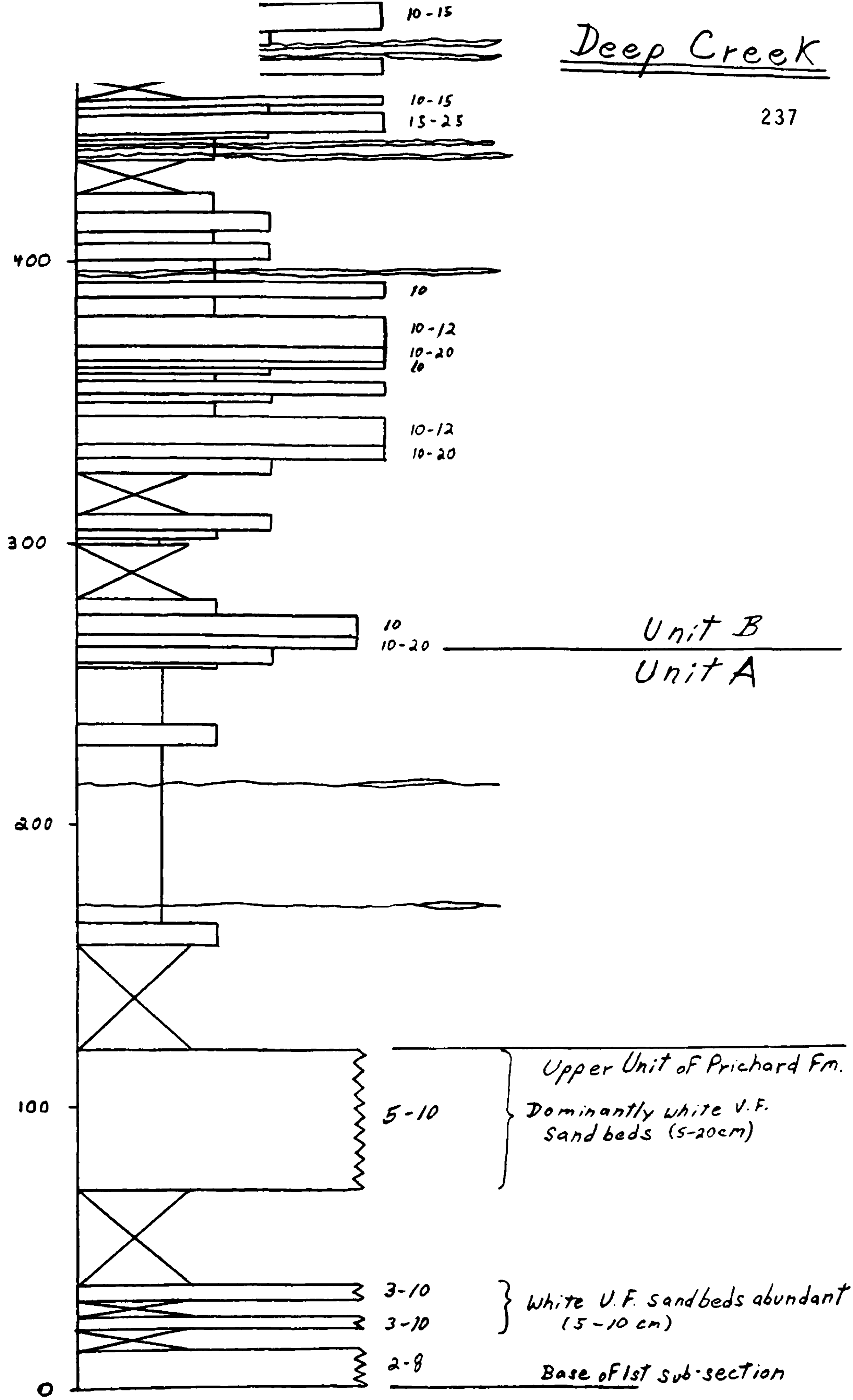
5900
5800
5700
5600
5500
ST. Regis correlative
Grinnell Fm
Sub-Types 1+3 mudchip
Conglomerates as in
5000'-5030'


6400
Sub-types 1+3 mudchip
conglom. as in 5000'-5030'
6300
Base of 4th sub-section.
St. Regis Correlative
Grinnell Fm.
Top of 3rd Sub-section.
6200
6100
30 (Single bed)
6000

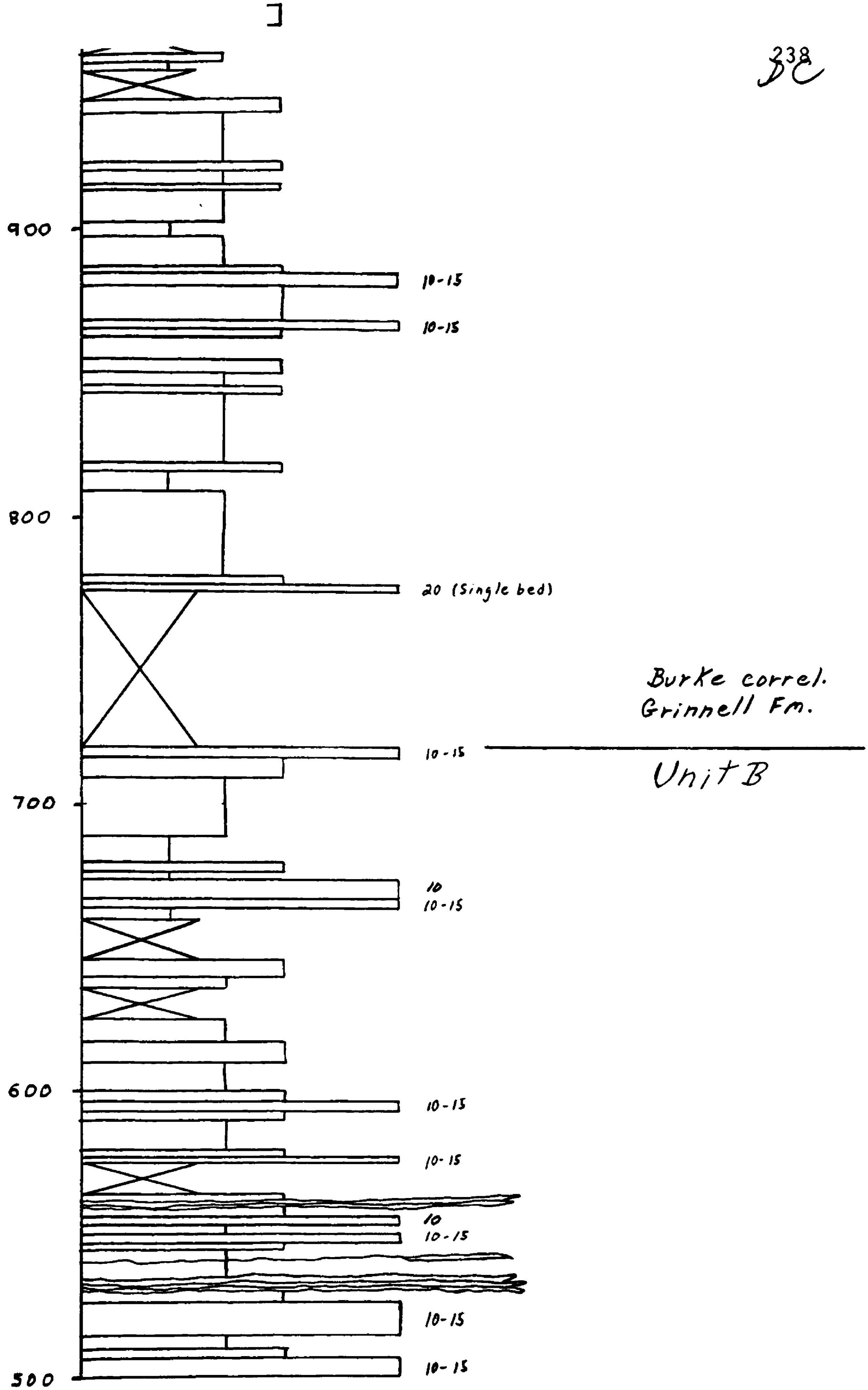

6900
6800
6700
6600
6500
Empire Fm.
St. Regis correl.
Grinnell Fm.

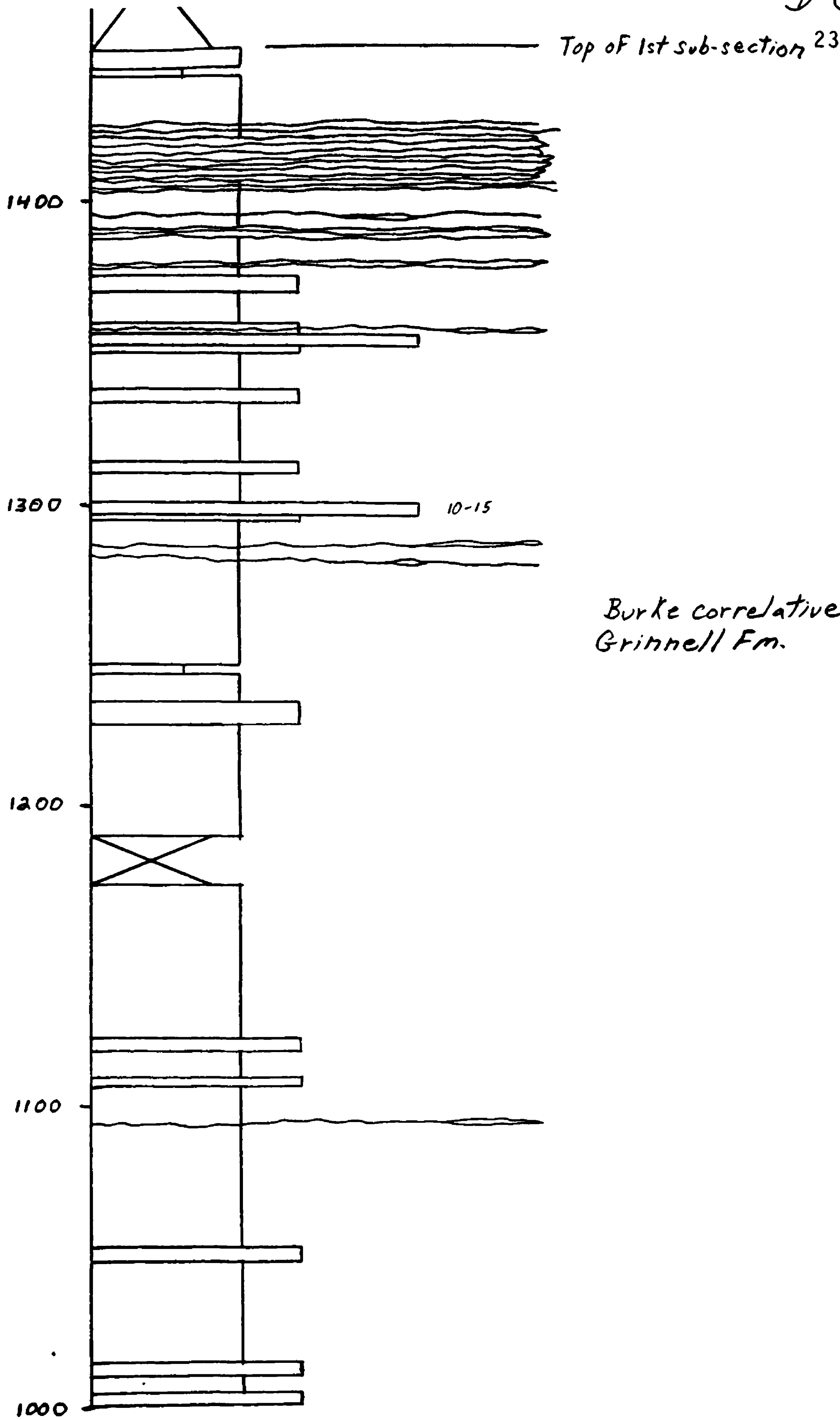

7300
Top of 4th sub-section
(End of outcrop)
Empire Fm.
7200
Interbeds of sub-type 2 mudchip conglom. dominate intervals up to 1 meter thick
(these low outcrops are just off the hiking trail.)
7100
7000

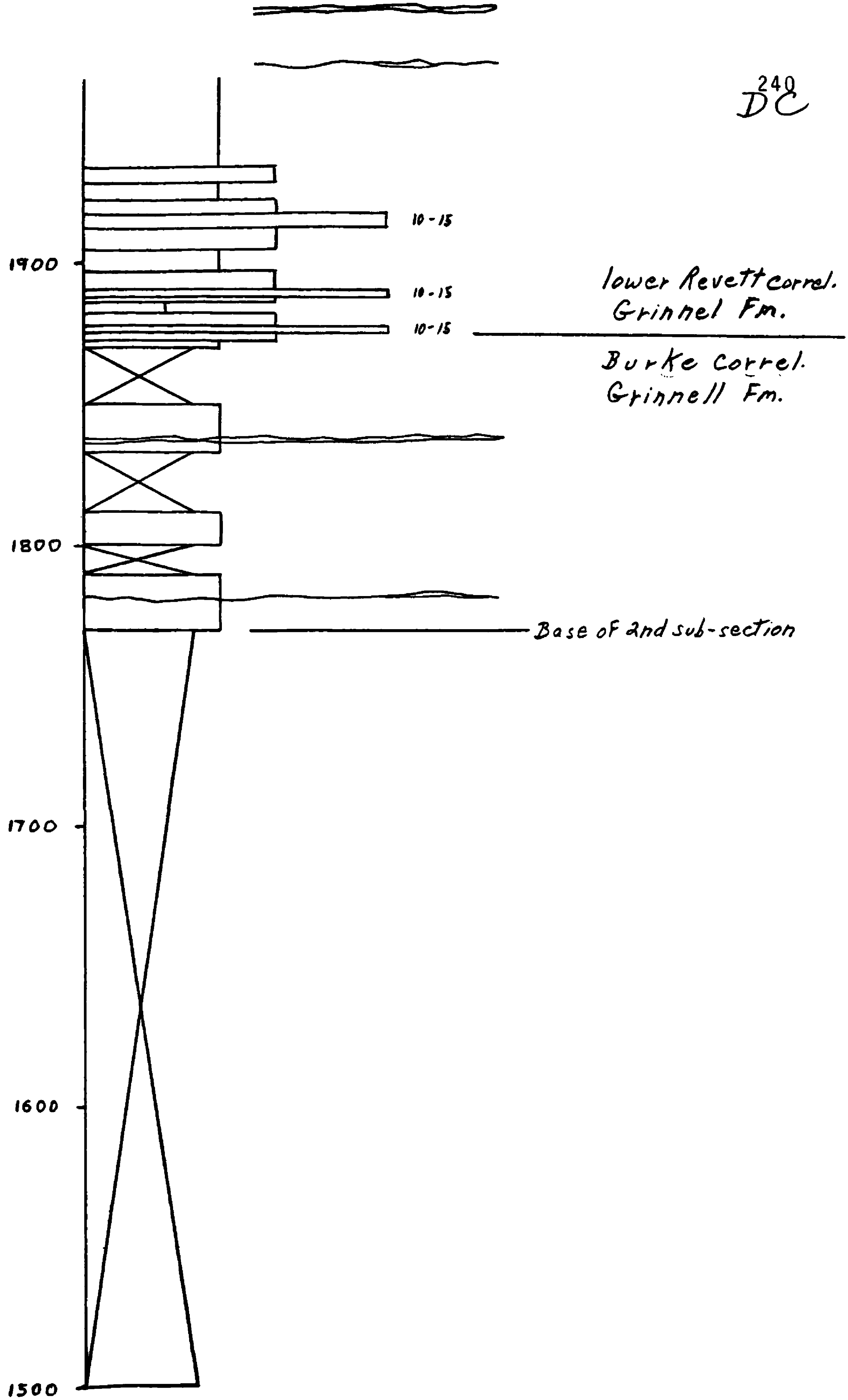
Deep Creek



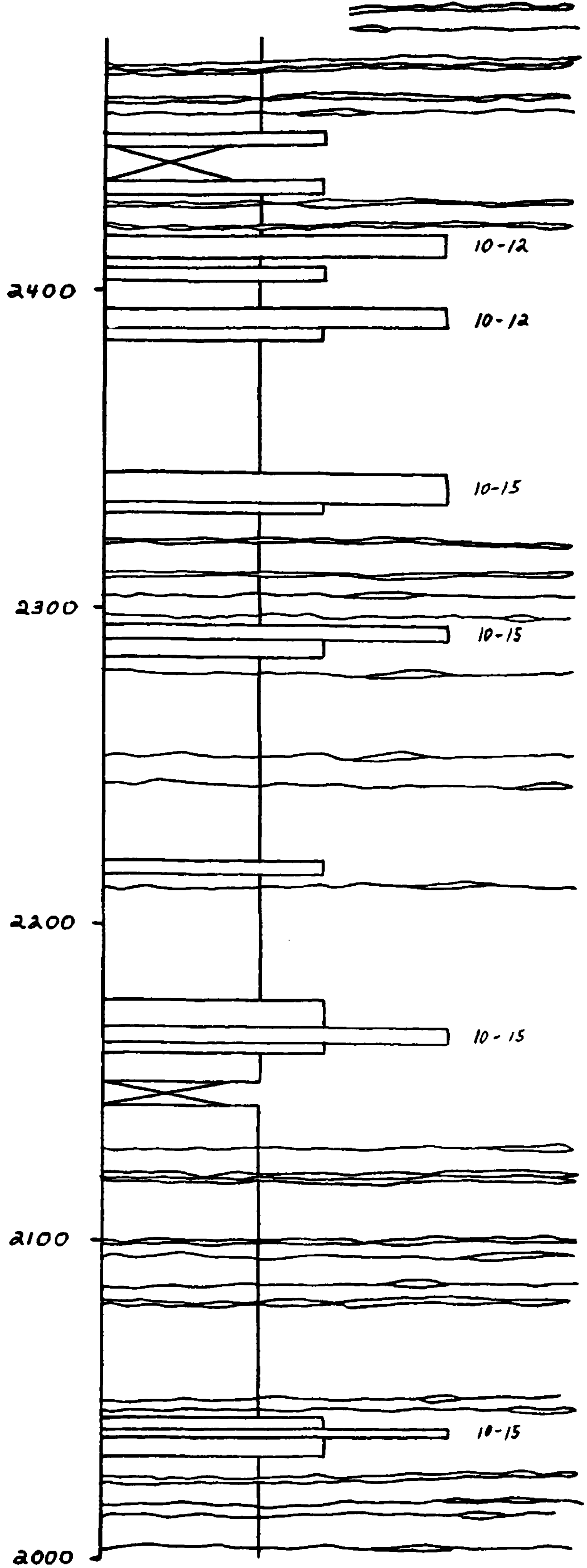
10-15
10-15
15-25
400
10
10-12
10-20
10
10-12
10-20
300
10
10-20
Unit B
Unit A
200
100
5-10
Upper Unit of Prichard Fm.
Dominantly white V.F. Sandbeds (5-20cm)
3-10
3-10
White U.F. sandbeds abundant (5-10 cm)
2-8
Base of 1st sub-section
0


900
800
700
600
500
10-15
10-15
20 (Single bed)
10-15
10
10-15
10-15
10-15
10
10-15
10-15
10-15
Burke correl.
Grinnell Fm.
Unit B


Top of 1st sub-section
1400
1300
10-15
Burke correlative
Grinnell Fm.
1200
1100
1000

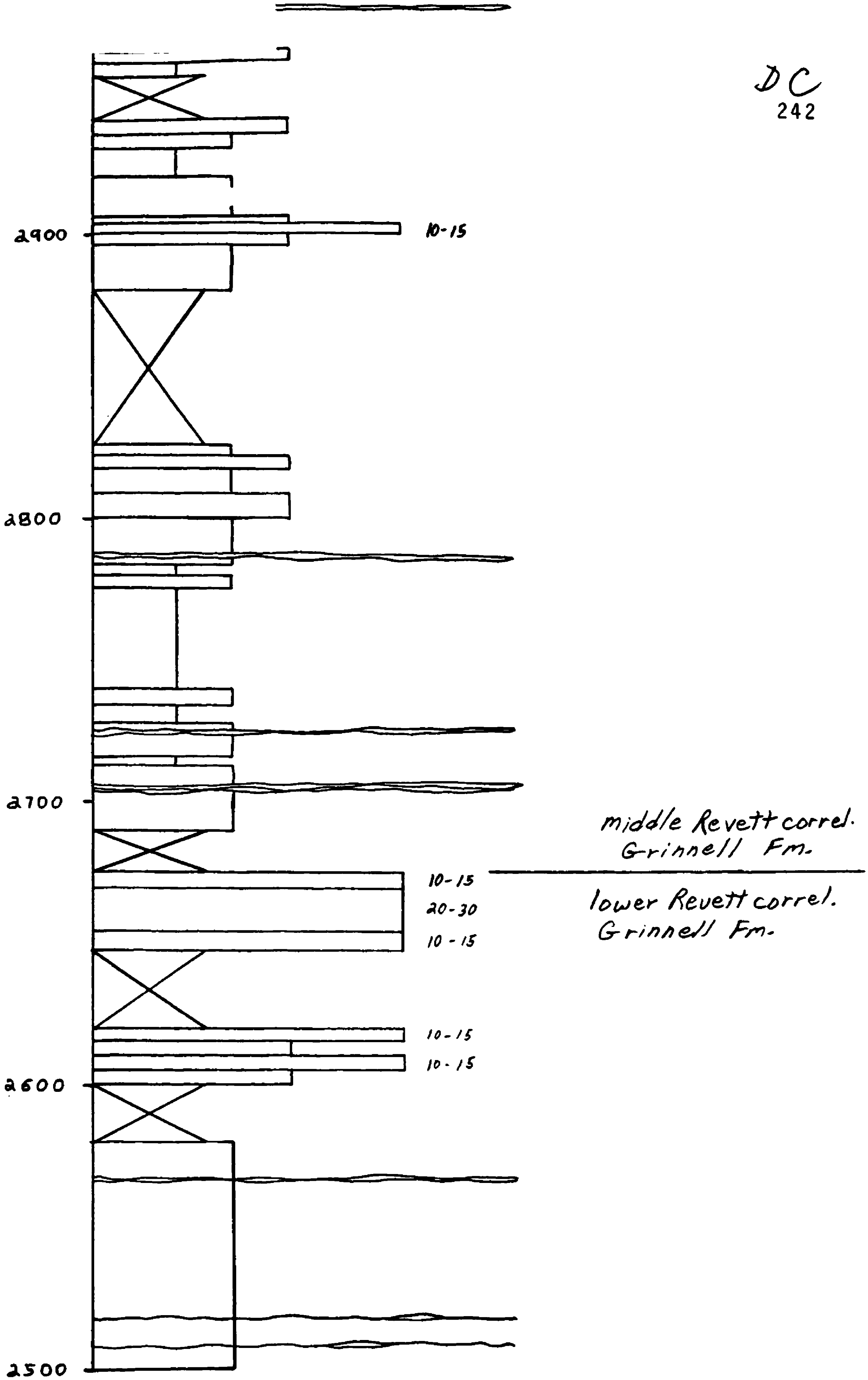

1900
1800
1700
1600
1500
10-15
10-15
10-15
lower Revett correl.
Grinnel Fm.
Burke Correl.
Grinnell Fm.
Base of 2nd sub-section

DC
241

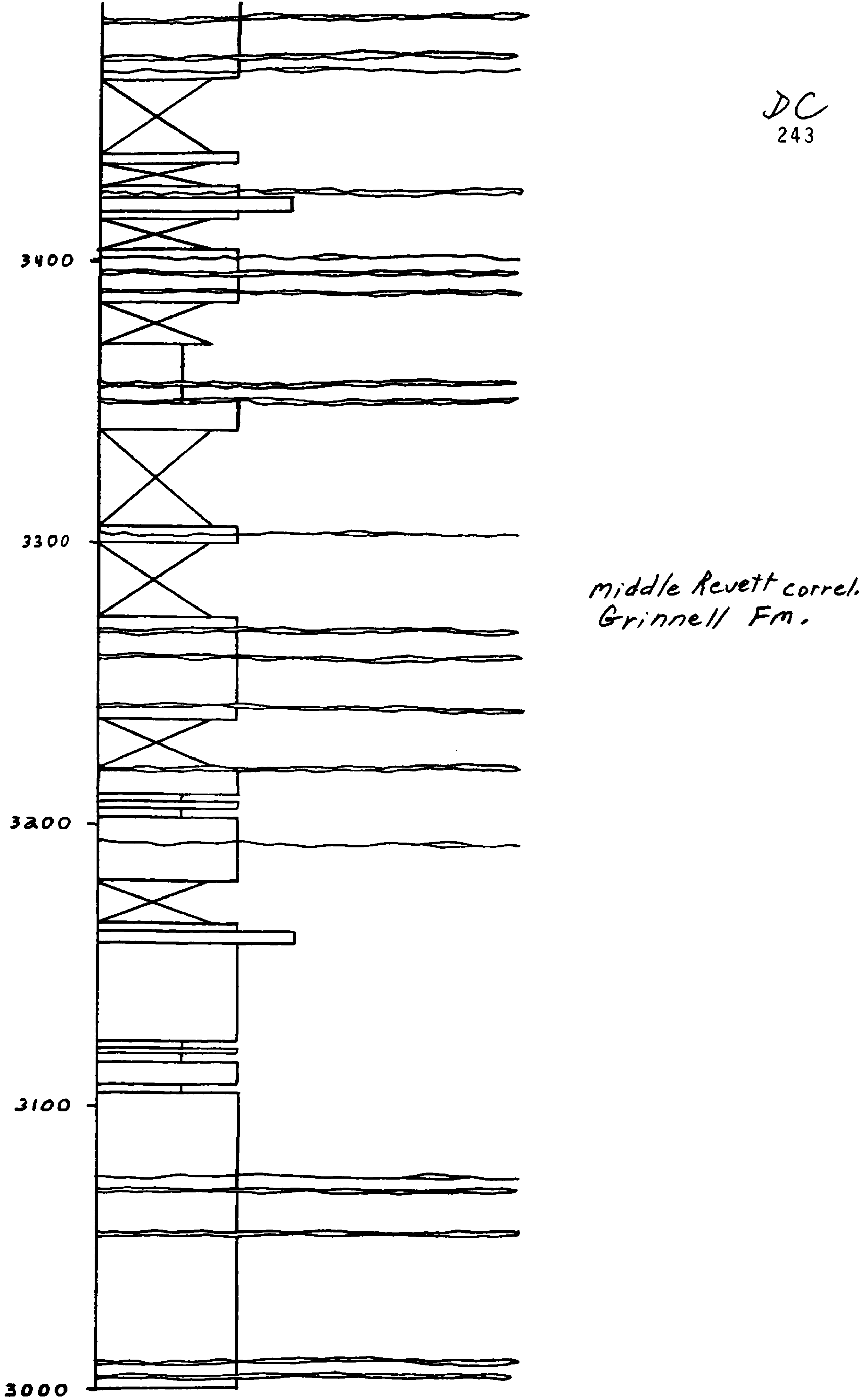


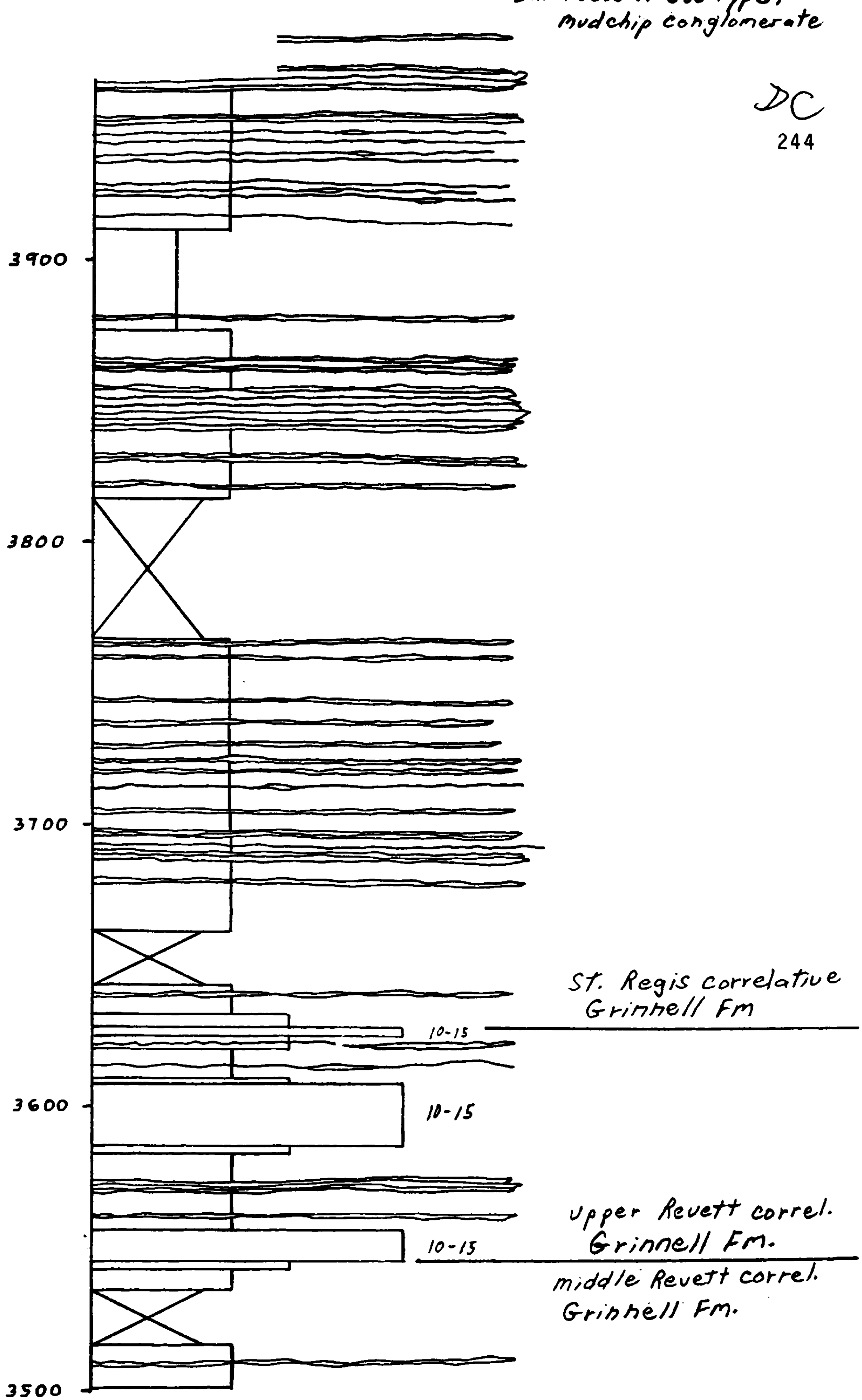
lower Revett correl.
Grinnell Fm.

DC
242

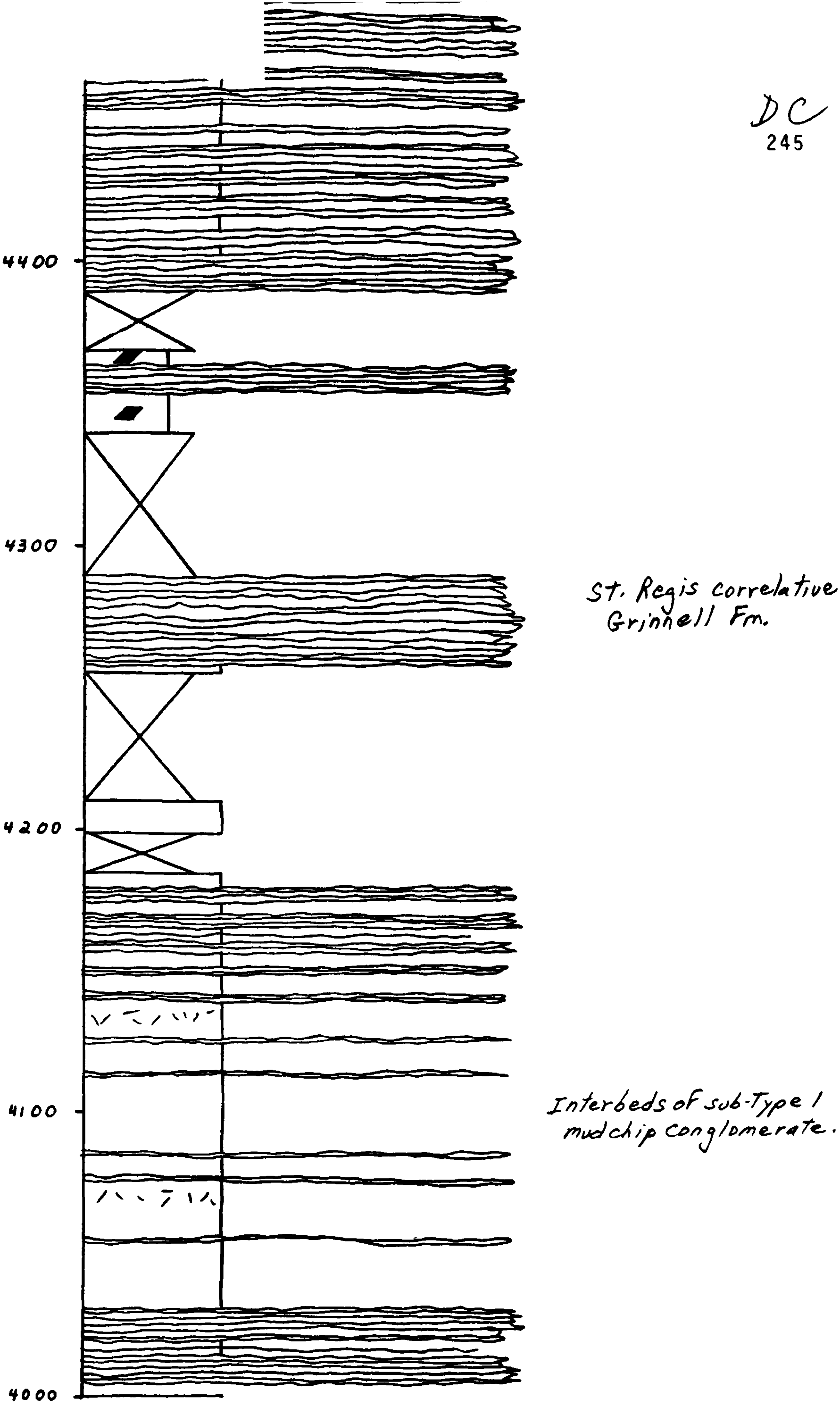

2900
10-15
2800
2700
middle Revett correl.
Grinnell Fm.
10-15
20-30
10-15
lower Revett correl.
Grinnell Fm.
10-15
10-15
2600
2500

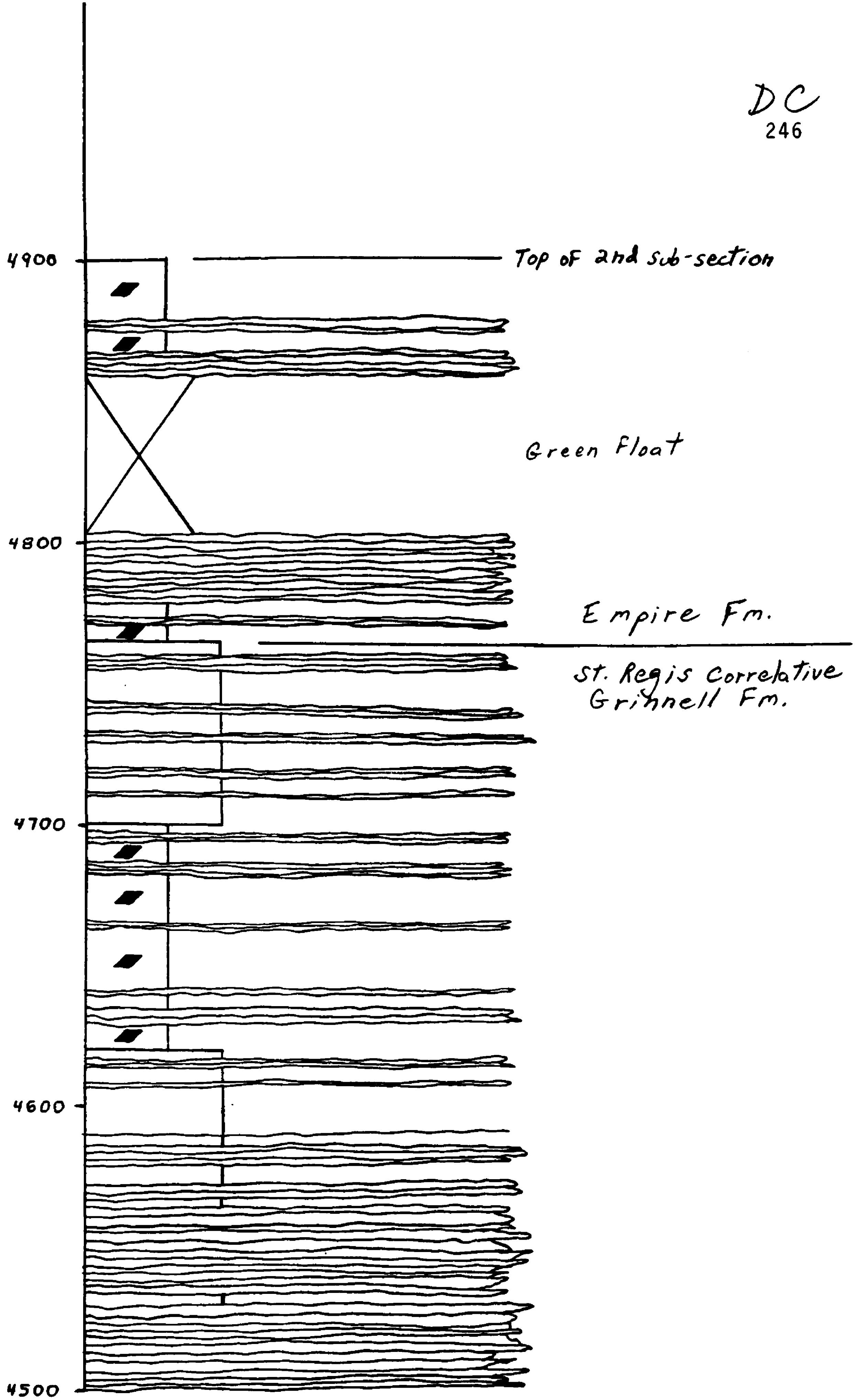
DC
243




mudchip conglomerate
3900
3800
3700
St. Regis correlative
Grinnell Fm
10-15
3600
10-15
upper Revett correl.
Grinnell Fm.
10-15
middle Revett correl.
Grinnell Fm.
3500

DC
245


4400
4300
4200
4100
4000
St. Regis correlative
Grinnell Fm.
Interbeds of sub-Type 1
mudchip conglomerate.


4900
Top of 2nd sub-section
Green Float
4800
Empire Fm.
St. Regis Correlative
Grinnell Fm.
4700
4600
4500