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**DETAILED SEDIMENTOLOGY OF THE BOWSER LAKE GROUP,
NORTHERN BOWSER BASIN,
NORTH-CENTRAL BRITISH COLUMBIA**

by

George Meredith Green, B. Sc. (Hons.)

**A thesis submitted to
the Faculty of Graduate Studies and Research
in partial fulfilment of
the requirements for the degree of
Master of Science**

**Ottawa-Carleton Geoscience Centre
and
Department of Earth Sciences
Carleton University
Ottawa, Ontario
September 1992**

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submitted by
George Meredith Green, B. Sc. (Hons.)
in partial fulfilment of the requirements
for the degree of Master of Science

Thesis Supervisor

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ABSTRACT

The Bowser Lake Group is a thick marine to non-marine succession within the Intermontane Belt of the Canadian Cordillera that records the syn- to late-orogenic interaction between volcanic rocks of Stikinia and oceanic rocks of the Cache Creek Terrane. In northern Bowser Basin Middle to Late Jurassic (Late Bathonian to Middle and possibly Late Oxfordian) rocks define an overall coarsening-upward/shoaling-upward sequence. The lower part (Ashman Formation) comprises three environmental assemblages: submarine fan (upper fan facies with poorly sorted, locally inverse graded pebble to boulder conglomerate); submarine canyon/gully (channel facies, with moderate- to well-sorted pebble conglomerate displaying inverse to normal grading and flow-parallel A-axis clast imbrication); and prodelta-slope (slump facies with internally deformed megablocks; overbank facies with thin- to medium-bedded fine- to medium-grained sandstone; slope facies with thinly laminated to bedded mudstone to very fine-grained sandstone). The middle part ("sandy lithosome") represents a shelf assemblage with facies that include: the outer shelf (thinly laminated to bedded siltstone to very fine-grained sandstone); the inner shelf (fine- to medium-grained sandstone with HCS and ripple cross stratification) and the nearshore (rusty weathering, medium-grained sandstone with parallel stratification and planar cross-stratification). The upper part ("rusty lithosome") comprises a fan-delta assemblage (braid-delta facies with planar and trough cross-stratified pebble to cobble conglomerates; fan/braid-delta slope (?) facies with massive, poorly sorted conglomerates; and fan delta front with erosively based parallel and planar cross-stratified conglomerates) and a coastal assemblage (gravel barrier-lagoon facies with planar and cross-stratified conglomerates displaying A/B clast imbrication).

Paleocurrent data (clast imbrication and cross-bedding) and clast compositions (predominantly grey, black and green radiolarian chert throughout the section) suggest that the most likely source area was the Cache Creek Terrane to the north/northeast. Gravels probably funnelled from fan- and braid-deltas (via sediment-gravity flows) to small canyons and gullies (in part old slump scars) across a narrow (2-4 km) shelf.

Meter-scale coarsening-upward cycles (consisting of siltstone to medium-grained sandstone) within the shelf assemblage probably was the result of some combination of eustatic sea-level changes, changes in basin subsidence and sedimentation rates associated with thrust loading of Cache Creek Terrane onto Stikinia, basement to the Bowser Basin. The overall coarsening-upward and shoaling of the Bowser Lake Group may be due to high sedimentation rates which counteracted the long term global sea level rise during this particular time period.

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Tony MacPhee assisted in the preparation of the maximum clast size versus bed thickness graphs. Jean Morin introduced me to McIntosh computers which enabled me (with much help) to create Figure 2. Kim Nguyen redrafted the post-defence version. Thanks Kim.

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Chapter 1 Introduction

1.1 Purpose and method of study

The Bowser Basin is a thick accumulation of syn- to late-orogenic, marine to nonmarine sedimentary rocks within the Intermontane Belt of the Canadian Cordillera (Eisbacher, 1974,b,c, 1981, 1985; Eisbacher, *et al.*, 1974). Recent sedimentological investigations of the Bowser Basin suggest that it developed as a foredeep formed by thrust loading of the Cache Creek Terrane over Stikinia, and thus preserves an indirect record of this tectonic process (Eisbacher, 1985; Ricketts and Evenchick, 1991).

As part of a regional Frontier Geoscience Program designed to place better constraints on the evolution of the Bowser Basin, a depositional model of the Bowser Lake Group has been developed by Ricketts (1990) and Ricketts and Evenchick (1991) for the northern part of the basin. However, detailed sedimentological studies were lacking until completion of the present study, based on six stratigraphic sections measured through the Ashman Formation and the overlying undivided strata of the Bowser Lake Group near the northern edge of the basin. Sections were measured at a scale of 1:400 and a total of 6400 m of section was measured. The location was chosen in the 1989 field season during regional mapping, and data were collected during the 1990 field season. Macrofossils were collected to constrain the age of the strata.

This thesis is divided into six chapters. Chapter 1 includes an introduction

to the study area and a short discussion of previous studies of Bowser Lake Group and Bowser Basin. Chapter 2 discusses the main regional elements associated with the present investigation, including an outline of the stratigraphy. Chapter 3 briefly summarizes the structural geology, and Chapter 4 provides detailed descriptions and interpretations of the lithofacies. Chapter 5 presents the results of a provenance study and Chapter 6 integrates the lithofacies observations in the form of a depositional model, and includes a discussion of some of its salient aspects.

The purpose of this thesis is to: (1) present detailed descriptions and interpretations of lithofacies, (2) provide paleocurrent data from imbrication studies of conglomerates, (3) present vertical and lateral clast composition trends within the section, and (4) integrate the above information to develop a detailed sedimentological model that augments the more regional studies of Ricketts (1990) and Ricketts and Evenchick (1991).

1.2 Location, access and topography

The study area is in the Skeena mountains, approximately 20 kilometres south of Tatogga Lake, north-central British Columbia (Figure 1). The area includes Tsatia Mountain and the western part of Todagin Mountain. Both Tsatia and Todagin mountains are accessible by helicopter or by a trail which follows Todagin Creek from Highway 37.

Figure 1.

Simplified map of the Canadian Cordillera showing the positions of the five morphogeological belts (see inset) in addition to the accreted terranes of the Intermontane Belt. Modified after Evenchick (1991c). The black box at the north end of the Bowser Basin shows the location of the study area and figures 3, 4, and 8.

Valley floor elevations average 925 m (3000 feet); mountain summits range from 2000 to 2350 m (6500 to 7800 feet). Slopes above tree line at approximately 1400 m (4600 feet) afford nearly 100% bedrock exposure. Volcanic and coarse clastic rocks (conglomerate) tend to produce rugged peaks and steep slopes; fine-grained clastic rocks tend to produce rounded peaks and moderate slopes, providing easy access along ridges.

1.3 Bowser Basin/Bowser Lake Group: previous work

The areal extent of the Bowser Basin was first delineated in 1956, during a Geological Survey of Canada helicopter reconnaissance mapping project designated "Operation Stikine" (G.S.C., 1957). During this project, the name Bowser Group was informally applied to an Upper Jurassic to Lower Cretaceous succession of shale, greywacke, and conglomerate in the vicinity of Bowser Lake (map 104 A/5, A/6-1:50 000). Duffell and Souther (1964) applied the same name to a similar aged sequence of marine to freshwater shale, argillite, greywacke, sandstone, conglomerate, and tuff in the northern third of the Terrace map-area (map 103 I, east half-1:250 000), approximately 200 km south of the Bowser Lake region. Souther (1972) continued to apply Bowser Group terminology in his geological report of the Telegraph Creek map-area (map 104G-1:250 000), which includes a small area in the northwestern part of the basin.

A tectonic synthesis of the north-central belt of the Cordillera, including the evolution of the Bowser Basin, was presented by Souther and Armstrong

(1966), who referred to lithologically and temporally similar strata as the Bowser Assemblage. Eisbacher (1974a,b,c), and Eisbacher *et al.* (1974) referred to the Bowser Assemblage in discussions of late-orogenic basins in the Canadian Cordillera.

Based on their work in the southern Bowser Basin, Tipper and Richards (1976) updated and redefined the stratigraphic nomenclature for Jurassic rocks in north-central British Columbia. They proposed the name Bowser Lake Group to replace the Bowser Group and Assemblage terminologies in recognition of the geographic feature from which the name originated. They also defined the Ashman Formation as the lowest formation within the Bowser Lake Group. Gabrielse and Tipper (1984) later assigned the Ashman Formation to lithologically similar strata in the northern part of the basin.

In the southern Bowser Basin, part of the definition of the Bowser Lake Group includes the elimination of Cretaceous strata, thereby restricting the age to Middle-Upper Jurassic (Tipper and Richards, 1976). This restriction is based on the following criteria: (1) lithological differences between Jurassic and Cretaceous strata (detrital muscovite occurs in Cretaceous but not in Jurassic strata); (2) a hiatus is recognized between Jurassic and Cretaceous strata; and (3) the Cretaceous basin of deposition is dissimilar to the Jurassic Bowser Basin (Tipper and Richards, 1976). The name Skeena Group has been applied to the remaining sedimentary and volcanic strata deposited during the Cretaceous Period (Hauterivian to Albian) (Tipper and Richards, 1976). In the northern half of the

Bowser Basin, the Bowser Lake Group is reported to range in age from Bathonian to Aptian (?) or Cenomanian (?) (Evenchick, 1991c), thereby extending its age into the Cretaceous.

Discovery of coal in the Groundhog coalfields, central Bowser Basin, has generated numerous studies of the stratigraphic, paleoenvironmental, tectonic, and economic characteristics of such deposits (Malloch, 1914; Buckham and Latour, 1950; Richards and Gilchrist, 1979; Bustin and Moffat, 1983; Bustin, 1984; Koo, 1986; Cookenboo and Bustin, 1989).

In the Groundhog coalfield area, the Currier, McEvoy, and Devil's Claw formations were formally proposed by Cookenboo and Bustin (1989) in conjunction with the palynological studies of Moffat *et al.* (1988); the first formation was assigned to the Bowser Lake Group and the latter two were assigned to the Skeena Group. The Currier, McEvoy, and Ashman Formation were also recognized outside the coalfields by Cookenboo and Bustin (1990). On the basis of a detailed sedimentological and paleontological study, Macleod and Hills (1990) suggested that all of the above formations proposed by Cookenboo and Bustin (1989) belong to the Bowser Lake Group. This was based on a sedimentological model of continuous sedimentation through the Jurassic-Cretaceous boundary and on biostratigraphic evidence, primarily invertebrate and plant macrofossils.

A comprehensive mapping program, named the Bowser Basin project, was initiated in 1985 to develop a regional structural and stratigraphic framework for

the northern part of the basin (Evenchick, 1986, 1987, 1988, 1989, 1991a,b,c; Evenchick and Green, 1990). The aim of that project was to provide a better understanding of the basin's evolution. Extensions of that project include biostratigraphic studies (Poulton *et al.*, 1991) and sedimentological analyses (Ricketts, 1990; Green, 1991; Ricketts and Evenchick, 1991). In addition, a study of the relationship between the Bowser Basin and its basement rocks is currently being conducted along the western flank of the basin (Greig, 1991, 1992). Other studies of basement rocks in the northern part of the basin include those of Smith *et al.* (1984), Thomson *et al.* (1986), Thorkelson (1988), and Marsden and Thorkelson (1992).

Chapter 2 Regional Geology

2.1 Introduction

The Bowser Basin is a syn- to late-orogenic sedimentary basin which preserves a record of terrane interactions within the Intermontane superterrane of the Canadian Cordillera. The main regional elements associated with the area of study include (1) the Intermontane Superterrane, (2) Stikinia, (3) the Cache Creek Terrane, (4) the Bowser Basin, and (5) the Bowser Lake overlap assemblage.

2.2 Regional Elements

2.2.1 Intermontane Superterrane

The Intermontane Superterrane is one of the five morphogeological belts of the Canadian Cordillera (Figure 1). Superterranes or composite terranes are composed of two or more independent terranes which were amalgamated before docking onto the western margin of North America (Gabrielse and Yorath, 1989). The Intermontane superterrane (Figure 1) is composed mainly of four fault-bounded terranes (Slide Mountain, Quesnel, Cache Creek, and Stikinia) which form a collage of marine to nonmarine volcanic and sedimentary rocks, and comagmatic granitic intrusions (Monger, 1984). Structural and stratigraphic evidence suggest that the terranes were amalgamated prior to Middle Jurassic accretion to the North American craton (Monger *et al.*, 1982; Monger, 1984; Monger, 1989).

2.2.2 Stikinia

Stikinia, the largest terrane in the Intermontane Belt, reaches a maximum width of approximately 300 kilometres in northern British Columbia (Figure 1). It is underlain by Devonian to Lower Jurassic felsic to mafic volcanic, volcanoclastic, carbonate, epiclastic, and minor plutonic rocks (Monger, 1984; Gabrielse, 1991). This terrane is considered to partially represent Late Paleozoic to Mesozoic island arc and back-arc environments (Tipper and Richards, 1976; Marsden and Thorkelson, 1992). Faunal (Monger and Ross, 1971; Smith and Tipper, 1986) and paleomagnetic evidence (Irving *et al.*, 1980; Monger and Irving, 1980; Irving and Thorkelson, 1990) suggest that Stikinia was farther south before its accretion to the North American craton.

The tectonic history of Stikinia is complex and protracted. In the northwestern part of Stikinia, Souther (1971) recognized Middle to Upper Triassic deformation characterized by uplift, tight folding, metamorphism, and intrusions. This episode, named the Tahltanian Orogeny, has not been studied in detail, and remains somewhat enigmatic. Uplift roughly coeval with deposition of the Hazelton Group in the Lower to Middle Jurassic produced the Stikine and Skeena arches (Figure 1)(Tipper and Richards, 1976; Marsden and Thorkelson, 1992). Souther and Armstrong (1966) suggested that the Stikine Arch may have been emergent by Late Triassic. The Skeena Arch was not fully emergent until Bajocian time (Tipper and Richards, 1976). Contractional deformation affected a broad area of northern Stikinia in the Late Jurassic(?) to mid-Cretaceous and

Late Cretaceous or Early Tertiary (Evenchick, 1991b). This is evident in the Skeena Fold Belt, a northwesterly trending feature characterized by various fold styles, thrust faults, subgreenschist metamorphism, and a frontal triangle zone (Evenchick, 1991b). The fold belt, estimated to represent at least 160 kilometres of crustal shortening, spans the width of the Bowser Basin and continues to the east side of the Coast Plutonic Complex (Evenchick, 1991b). It affects both volcanic rocks of Stikinia (the basement to the Bowser Basin) as well as the Bowser Lake and Sustut Groups.

2.2.3 Cache Creek

The Cache Creek Terrane extends along most of the eastern side of Stikinia (Figure 1). In north-central British Columbia, Cache Creek was structurally placed on top of Stikinia by the north/northeast-dipping King Salmon Fault (Figure 1). Deformation and overthrusting of the northern Cache Creek along the King Salmon Fault is constrained to Bajocian and older (Gabrielse, 1991).

The Cache Creek Terrane, which contains a wide variety of lithologies including ultramafic and carbonate rocks, argillite, tholeiitic to basaltic rocks, and radiolarian-bearing chert (Monger *et al.*, 1982; Monger, 1984; Gabrielse, 1991), is interpreted to represent an oceanic environment with associated volcanic arcs. Shallow-water carbonates in Cache Creek strata may represent atolls deposited on seamounts (Monger, 1989) or intra-arc or forearc environments (Gabrielse, 1991).

Fossil evidence suggests that the Cache Creek oceanic environment ranged in age from Mississippian to Late Triassic (Coney *et al.*, 1982; Monger *et al.*, 1982; Monger, 1984; Gabrielse, 1991). However, ribbon chert in the northern Cache Creek of southern Yukon contain radiolaria ranging from Late Carnian to Middle Norian, Late Norian (Late Triassic) to Post-Hettangian; and possibly Late Sinemurian or perhaps Pliensbachian to early Toarcian (Early Jurassic) (Cordey *et al.*, 1991). In southern British Columbia, radiolaria in strata of the western belt of the Cache Creek provide an upper age limit of Pliensbachian to Bajocian (Cordey *et al.*, 1987). Also, a chert clast from the Bowser Lake Group in the northern Bowser Basin contained radiolaria of Early Jurassic age (F. Cordey, pers. comm., 1992). This suggests that the Cache Creek Ocean was at least partially open up to and including the Early Jurassic. This places doubts on the suggestion that superterrane I (Intermontane) defined by Monger *et al.* (1982) may have been assembled by the Late Triassic (Cordey *et al.*, 1987).

2.2.4 Bowser Basin

The Bowser Basin covers approximately 50000 km² of Stikinia in north-central British Columbia (Figure 1). The basin consists of a thick (greater than 5000 metres) accumulation of marine to nonmarine clastic rocks of the Bowser Lake and Spatsizi Groups. It is bounded on the east by the Sustut Basin, on the southwest by the Coast Plutonic Complex, and on the west, north, and northwest by volcanic and volcanoclastic rocks of Stikinia.

Various hypotheses deal with the tectonic setting of the Bowser Basin of deposition. Souther and Armstrong (1966) suggested that deposition occurred in a structural depression which was partially open to the west, bounded by high angle faults, and to the north and south by the Stikine and Skeena arches respectively.

Eisbacher (1974a,b,c, 1981; Eisbacher *et al.*, 1974) labelled the Bowser Basin a successor basin: a trough or depression forming on deformed and intruded Stikinian volcanic rocks. Suggestions have been made that the Bowser Basin developed within the pre-existing Lower to Middle Jurassic Hazelton Trough (Tipper and Richards, 1976; Marsden and Thorkelson, 1992). The elliptical shape of the present-day basin may reflect the original outline of the Stikine arc (Eisbacher, 1981); however, as much as 160 kilometres of shortening has drastically altered the shape of the original basin (Evenchick, 1991b).

Foredeep models for Bowser sedimentation involve the loading of Stikinian crust by the Cache Creek Terrane along southwest-vergent thrusts (King Salmon and Nahlin Faults) (Wheeler and Gabrielse, 1972; Monger, 1977; Eisbacher, 1985; Gabrielse, 1991). The subsequent flexural response to loading of Stikinia may have produced the sink for oceanic detritus of the uplifted Cache Creek terrane from the north/northeast (Ricketts and Evenchick, 1991). Initiation of the Bowser Basin foredeep is postulated to have begun in Aalenian time (Middle Jurassic), with deposition of the Abou Formation of the Spatsizi Group (see next section)(Ricketts and Evenchick, 1991). This transgressive sequence overlies a

disconformable surface within the Spatsizi Group and may represent the starved basin phase of a developing foredeep (Ricketts and Evenchick, 1991). Deposition of the first coarse clastics from the uplifted Cache Creek Terrane to the northeast did not occur until Bathonian time (Ricketts and Evenchick, 1991; Ricketts *et al.*, in press).

2.2.5 Bowser Lake Overlap Assemblage

The primary source of detritus (predominantly chert) for the Bowser Lake Group is the Cache Creek Terrane to the northeast (Souther and Armstrong, 1966; Tipper and Richards, 1976; Eisbacher, 1981; Currie, 1984). The Bowser Lake Group thus records obduction of the Cache Creek terrane onto Stikinia, although it does not overlie the boundary between the two terranes, and it is therefore considered an overlap assemblage on Stikinia called Bowser Lake (Wheeler *et al.*, 1991).

2.3 Stratigraphy of the study area

Three Mesozoic stratigraphic units are regionally mappable in and around the northern part of the study area; these are the Hazelton, Spatsizi, and Bowser Lake Groups (Figure 2). The informally named Maitland volcanics constitute a fourth and youngest (Early Pliocene) stratigraphic unit immediately around the study area (Figure 3).

Figure 2.

Mesozoic stratigraphic units in and around the study area, northern Bowser Basin. The schematic section on the right-hand side summarizes the overall coarsening-upward trend, approximate age range (Late Bathonian to Middle and possibly Late Oxfordian), and relative vertical positioning of assemblages observed within the study area. The vertical lines between the two columns indicate the assemblages covered by each measured stratigraphic section (section identified by Roman Numeral). Line lengths are not to scale. Modified after Evenchick (1991c).

Figure 3.

Generalized geology map of the study area, showing the gross stratigraphy, location of measured stratigraphic sections (EPG-I-VI), and axial surface traces of major folds. Most geological contacts and northeasterly trending axial surface traces are adapted from Evenchick and Green (1990). The map on the right represents the outline of the Bowser Basin.

2.3.1 Hazelton Group

Most volcanic rocks surrounding the Bowser Basin are assigned to the Hazelton Group, a term that refers to Early to Middle Jurassic (Sinemurian to Lower Bajocian (H.W. Tipper, pers. comm. to D.J. Thorkelson)), "... basaltic to rhyolitic volcanic rocks, sedimentary rocks, their tuffaceous equivalents, and minor limestone..." (Tipper and Richards, 1976). Marsden and Thorkelson (1992) expanded on the above description, suggesting that the Hazelton Group include all Lower to Middle Jurassic volcanic and related sedimentary rocks deposited on Stikinia south of about 58° N latitude. This would include, for example, the Lower Pliensbachian to Lower Bajocian Spatsizi Group defined by Thomson *et al.* (1986) in the Spatsizi map area. Detailed descriptions of the various Hazelton Group formations, members, and facies delineated in the area southeast of the Bowser Basin are provided by Tipper and Richards (1976). The Hazelton Group has been interpreted to represent island arc volcanism and associated clastic deposition in the Hazelton Trough (Tipper and Richards, 1976). The trough may have formed in an interarc or back-arc environment (Tipper and Richards, 1976), possibly as a region of back-arc extension between two volcanic arcs (Marsden and Thorkelson, 1992).

In the northern part of the Bowser Basin (Spatsizi map area 104H), strata of the Hazelton Group are informally known as the Cold Fish (Evenchick, 1986; Thomson *et al.*, 1986) and Brock volcanics (Read and Psutka, 1990). They are composed of felsic to mafic lava flows and tuffs; most are subaerial, but shallow

marine successions are also present (Thorkelson, 1988). Ages of the Cold Fish and Brock volcanics are Early Pliensbachian and Early Toarcian, respectively (Thorkelson, 1988; Read and Psutka, 1990).

The volcanic rocks in western Spatsizi, closest to the study area, were examined by Evenchick and Green (1990). These rocks consist of rhyolitic to basaltic rocks including tuff, breccia, and volcanogenic sandstone to pebble conglomerate. This succession is reported to be Bajocian in age (Smith *et al.*, 1984), although the conglomerate closely resembles sedimentary rocks of Late Triassic to Early Jurassic age in central Spatsizi map area (Thorkelson, pers. comm., 1991). Pliensbachian fossils are also reported (H.W. Tipper, pers. comm. in Evenchick, 1991a)

2.3.2 Spatsizi Group

The Pliensbachian to Bajocian Spatsizi Group, which is best exposed in northern Spatsizi map area, is considered to be a coeval marine sedimentary succession of the Cold Fish volcanic pile (Smith *et al.*, 1984; Thomson *et al.*, 1986). Outcrops of similar strata in other parts of the Bowser Basin suggest that it may underlie all of the Bowser Lake Group (Thomson *et al.*, 1986).

Thomson *et al.* (1986) divided the Spatsizi Group into five formations; in ascending order, they include the Joan, Wolf Den, Melisson, Abou, and Quock formations. They are composed of varying amounts of tuffaceous shale, shale to boulder conglomerate, and limestone (Smith *et al.*, 1984; Thomson *et al.*, 1986).

The following descriptions and interpretations are based on the work of Thomson *et al.* (1986). The Joan Formation overlies the Cold Fish volcanics. A basal conglomerate with clasts ranging up to boulder size suggests an erosional contact with the underlying volcanic rocks. The formation generally fines upward to quartz siltstone, representing a transgressive event in a shallow to moderate water depth.

The Late Pliensbachian to Middle Toarcian Wolf Den Formation conformably overlies the Joan Formation. It consists predominantly of grey to black shale and represents a significant water depth increase over the Joan Formation; conditions are interpreted to have been quiet and anaerobic.

The conformable Melisson Formation is composed of calcareous to siliceous siltstone to sandstone; siliceous units are attributed to pyroclastic eruptions. Parallel and ripple laminations in sandstones and siltstones respectively, together with a relative coarsening over the previous formation, may represent a shoaling event during the Late Toarcian. Erosion related to shoaling may also account for missing sections of the Melisson Formation.

The poorly exposed Late (?) Toarcian to Aalenian Abou Formation lies unconformably over the Melisson Formation; it consists of grey weathered calcareous to siliceous (volcanogenic) shales with high (20%) organic contents. Normal grading is present in some fine beds. High organic contents, laminations, and the predominance of shales indicate the return of anaerobic to dysaerobic deep-water conditions after the shoaling associated with the Melisson Formation.

The Early Bajocian Quock Formation conformably overlies the Abou Formation. It is composed of alternating light and dark grey beds, giving the formation a striped appearance; hence, the informal name "pyjama beds" has also been applied to this formation. The light grey beds contain silt-sized quartz and plagioclase in a chert matrix. The darker beds contain relatively high amounts of clay and organic matter. Much of the formation has a rusty appearance due to the oxidation of pyrite. The dark coloured beds of the Quock Formation imply quiet anoxic conditions, with the deposition of radiolaria tests and clay-size particles. The lighter coloured beds include air fall tuffs and resedimented tuffs deposited by density currents related to volcanism to the north. Convolute bedding in the Quock Formation suggests unstable slope conditions, perhaps as a consequence of seismic and volcanic activity. The Quock Formation was observed at one locality near the base of section EPG-I (Plate 1). Overburden covers its contact with the overlying Bowser Lake Group.

The regional interpretation for the northern Spatsizi map area includes the development of a coeval sedimentary basin south of the Cold Fish volcanic arc. During the Pliensbachian to Early Bajocian periods, varying amounts of volcanic detritus accumulated in the basin. The dominant facies trends include a southward decrease in clastic size and volcanic input away from the volcanic arc (Smith *et al.*, 1984; Thomson *et al.*, 1986). A relatively major regressive event occurred during the Middle Toarcian which caused shoaling and partial erosion of the Melisson Formation. An Aalenian transgressive event produced the deep water conditions typical of the Abou and Quock formations (Thomson *et al.*, 1986). Cold Fish volcanism and associated deposition of the Spatsizi Group sediments is reported to have ceased in the Early Bajocian (Thomson *et al.*, 1986).

Plate 1.

View to the southeast of "pyjama beds" of the Quock Formation, approximately 50 m below section EPG-I.

2.3.3 Bowser Lake Group

Strata of the Spatsizi Group are overlain by a minimum of 4000 m of shale, siltstone, sandstone, and pebble to cobble conglomerate of the Bowser Lake Group (Evenchick, 1991b). Locally, the contact between the two groups is reported to be an unconformity (Monger, 1977). Thomson *et al.* (1986) described the contact as an angular discordance measuring 5°. However, on the eastern side of the Bowser Basin, the Bowser Lake Group conformably overlies the Spatsizi Group (Evenchick, 1991c). A similar relationship exists between the Bowser Lake Group and the underlying Salmon River Formation of the Hazelton Group on the west side of the basin (Anderson and Thorkelson, 1990).

The lowest part of the Bowser Lake Group consists of the regionally mappable Ashman Formation, a term first defined by Tipper and Richards (1976) in the southern Bowser Basin. In the northern part of the basin, the Ashman Formation composes 800 metres of the first stratigraphic section (EPG-I). Evenchick and Green (1990) reported a 1400 metre section of Ashman Formation strata north of Todagin Creek. It consists mainly of dark grey shales and siltstones monotonously interbedded with buff weathered mudstones to very fine sandstones (Plate 2a). Variable amounts of fine to medium grained sandstones and lenses of chert pebble to cobble conglomerates occur sporadically within the relatively incompetent fine grained strata (Evenchick, 1987; Green, 1991). Ashman Formation in the northern part of the Bowser Basin has also been

Plate 2.

- A. View to the west showing the typical striped appearance of the Ashman Formation. The black arrow points to an example of the overbank facies of the prodelta-slope assemblage.
- B. View to the south of ridge-forming maitland volcanic rocks unconformably overlying Bowser Lake Group strata (hidden under talus).
- C. Close-up of peculiar weathering pattern in sandstone of the shelf assemblage. The weathering pattern is due to a combination of joints, irregular fractures, and possibly spaced cleavage.
- D. View to the east of a relatively large offset in strata of the Ashman Formation. Scale indicated by figure (arrow).

documented by Currie (1984), Gabrielse and Tipper (1984), Evenchick (1986, 1989, 1991b), Ricketts (1990), and Ricketts and Evenchick (1991). The age of the Ashman Formation in the northern part of the basin is Late Bathonian to Early Oxfordian (Gabrielse and Tipper, 1984; Ricketts and Evenchick, 1991). A Late Bajocian to Early Oxfordian age is recorded in the southern part of the basin (Tipper and Richards, 1976).

A general transition from marine to nonmarine is documented for the remaining Bowser Lake Group strata (Eisbacher, 1974a,b,c; Eisbacher *et al.*, 1974; Tipper and Richards, 1976; Richards and Gilchrist, 1979; Eisbacher, 1981; Bustin and Moffat, 1983; Evenchick and Green, 1990). This, in concert with an overall coarsening upward trend, reflects rapid progradation and shoaling of the clastic succession (Eisbacher, 1981). The strata above the Ashman Formation have not been divided into formations in the northern part of the basin. This is due to the lack of regionally recognizable marker units along with the probable vertical and lateral repetition of the various lithofacies (Evenchick, 1988, Evenchick and Green, 1990). The formations defined by Cookenboo and Bustin (1989) in the north-central part of the basin, above the Ashman Formation, have not been recognized in and around the study area (Evenchick and Green, 1990). The youngest recorded age for the Bowser Lake Group in the northern Bowser Basin, immediately around the study area, is Late Oxfordian (Poulton *et al.*, 1991).

2.3.4 Maitland volcanic rocks

The youngest strata in the study area are the informally named Maitland volcanics (Plate 2b)(Souther, in press). They occur as isolated outcrops on the upper reaches of Tsatia Mountain (Figure 3) and lie unconformably on the Bowser Lake Group. The maximum flow thickness on Tsatia Mountain is approximately 30 m. Evenchick and Green (1990) gave a range in flow thickness of 3-350 m for units in the Spatsizi map area. This Late Pliocene volcanic suite is related to the Mount Edziza volcanic complex situated 50 km to the west (Souther, in press).

Chapter 3 Structural Geology

3.1 Introduction

Strata of the Bowser Lake Group in the northern part of the Bowser Basin are structurally disrupted by a complex array of folds and faults within a zone known as the Skeena Fold Belt (Evenchick, 1991b). Consequently, problems posed by the structural duplication of strata, and the juxtaposition of similar packages of strata representing highly disparate depositional environments, render the task of deciphering the local stratigraphy difficult. A detailed understanding of the regional structural style is essential to ensure that locally abundant soft-sediment structures such as load structures and slump folds are not mistaken for structures formed by tectonic processes.

In this study, an area of relatively simple structure was chosen to facilitate the documentation and interpretation of the stratigraphic record. The area was also chosen for its proximity to a known stratigraphic level: the contact between the Bowser Lake Group and the underlying Spatsizi Group. A brief overview of the Skeena Fold Belt as defined by Evenchick (1991b), supported by cross sections and stereogram, will be presented below to summarize the structural style of the study area.

3.2 Skeena Fold Belt

Reflecting major contractional deformation, the Skeena Fold Belt embraces a large part of northern Stikinia, including the Bowser Basin and part of the Sustut Basin to the northeast (Evenchick, 1991b; her Figure 1). The fold belt extends to the east side of the Coast Plutonic Complex; the southern boundary follows the most southerly occurrence of the Bowser Lake and Skeena groups (Figure 1)(Evenchick, 1991b). The fold belt comprises deformed strata of the Spatsizi, Bowser Lake, Sustut, and Skeena groups as well as Middle Jurassic and older volcanic and sedimentary rocks to the northwest, west, and south of the Bowser Basin.

The morphology of the Skeena Fold Belt is similar to that of the Rocky Mountain Fold and Thrust Belt to the east. Some of the main features associated with the Skeena Fold Belt include: an associated metamorphic/plutonic belt (Coast Plutonic Complex); a diverse range of fold styles and geometries in which axial surfaces predominantly trend northwest; an inferred eastward younging of structures away from the metamorphic/plutonic belt; an interpreted decollement(s) into which thrust faults sole; a deformed foreland basin (Sustut Basin); and a frontal triangle zone (Evenchick, 1991b).

An enormous amount of northeast crustal shortening is recorded within the Skeena Fold Belt. Evenchick (1991b) reported an average shortening of 44%; a range of 32% to 60% shortening was recorded for specific areas. These figures are minimum estimates and do not include, for example, penetrative deformation

which is documented in some of the fine-grained lithologies of the Bowser Lake and Spatsizi groups (Evenchick, 1991b). The average 44% crustal shortening translates to approximately 160 kilometres across the width of the belt. This implies that the 200-kilometre-wide Skeena Fold Belt, which incorporates the Bowser Basin and underlying Stikinia, was once 360 kilometres wide (Evenchick, 1991b).

The age of deformation for the Skeena Fold Belt is based on structural, sedimentological, and palynomorphic evidence. The oldest recorded deformation may have been as old as Oxfordian or as young as Albian. The youngest age for contractional structures is bracketed between latest Cretaceous to early Tertiary (Evenchick, 1991b). According to Evenchick (1991b), these time constraints, and the interpretation that structures root to the west, support the hypothesis that formation of the Skeena Fold Belt was closely related to evolution of the Coast Plutonic Complex.

3.3 Structural Style in the Study Area

The study area is near the northwestern edge of the Skeena Fold Belt, western Spatsizi map sheet (Figure 3). Crustal shortening in this part of the fold belt is more subdued than regions to the south and east, but is characterized by a wide variety of fold styles and orientations (Evenchick and Green, 1990). With the exception of one northwest-trending axial surface trace (Figure 3), the main trend of axial surface traces of folds is northeasterly. This contrasts with the

predominant northwesterly trend observed in other areas in the fold belt.

Folds in the study area are generally larger in scale and more gentle than other folds of the western Spatsizi map area (Evenchick and Green, 1990).

Figure 4 shows two representative structural cross sections. Together, sections AB and CD reveal that the folds are gentle, with rounded hinges, and display a basin and dome interference geometry, with no observable fold vergence. Based on measurements taken from the map and cross sections in Figure 4, the wavelength and amplitude of the folds are 2.3 km and 350 m, respectively. Figure 5 displays a stereogram summarizing the poles-to-bedding measured within the study area. The poorly defined best-fit great-circle girdle for the poles indicates that fold plunges are gentle.

As in the rest of the Skeena Fold Belt (Evenchick, 1991b), cleavage in the study area is mainly controlled by the lithology of the affected strata. Fine-grained rocks such as shales and siltstones of the Ashman Formation and overlying strata tend to exhibit a closely spaced penetrative cleavage that in most places has destroyed any primary sedimentary structures. Fine- to medium-grained sandstones display various degrees of spaced cleavage. For example, near the peak of Tsatia Mountain, high-angle spaced cleavage in association with irregular fractures and joints, produced a peculiar weathering pattern (Plate 2c). The conglomerates rarely display cleavage, as is readily observed in the Ashman Formation, where relatively pristine conglomerate units are surrounded by highly cleaved siltstones and shales.

Figure 4.

Structural cross-sections (AB, CD) showing the structural style within the study area. Locations of the cross-section traces are shown in the map at the top left of this figure.

Figure 5.

Equal area stereogram showing the poles-to-bedding obtained from stratigraphic sections EPG-I to VI, northern Bowser Basin.

Faults thwart efforts to obtain a coherent sedimentological reconstruction. Unlike areas to the east (East Spatsizi map sheet), where faults with significant amounts of displacement are relatively common, evidence of large-scale faults on Todagin and Tsatia mountains is lacking. One minor high-angle fault with subhorizontal slickensides was observed in a conglomerate unit near the base of section EPG-I (Todagin Mountain). The amount of displacement along this fault could not be established, although it does not appear to be significant.

Stratigraphically repeated orange-buff weathered units provide excellent markers to accurately define both contractional and extensional faults within the fine-grained strata of the Ashman Formation (Plate 2d). These displacements are commonly low angle and are characterized by sharp cutoffs. Because offsets are numerous and range from a few centimetres up to tens of metres, caution had to be exercised to avoid duplicate measurement of structurally repeated packages of strata. An areal survey revealed no evidence of larger offsets which might not have been detected during measurement of the sections. However, many parts of the lower section of EPG-III (northeast side of Tsatia Mountain) are covered by overburden, and so the possibility of structural duplication within this region cannot be ruled out. An additional problem arises as to whether to classify these faults as tectonic or sedimentological (of soft-sediment origin). This issue will be addressed when slump structures are discussed.

Todagin and Tsatia mountains are separated by a prominent valley parallel to the northeast-trending axial surface traces of folds. The possibility that a major

fault exists within this valley poses a significant problem for stratigraphic correlation. Approximately 10 kilometres east of the study area, this valley is terminated by a north- trending ridge system. Reconnaissance mapping along this ridge, as well as within the area of northeasterly-trending folds, has revealed no evidence of faulting. Consequently, no displacement of strata is assumed to have occurred between the two mountains. This renders sedimentological and stratigraphic correlations between Tsatia and Todagin mountains straightforward.

Chapter 4 Lithofacies

4.1 Introduction

Based on preliminary examination of the Middle to Upper Jurassic Bowser Lake Group, Ricketts (1990) proposed a depositional model for the northern part of the Bowser Basin. Rickett's model comprises a spectrum of environments including deep-sea submarine fans, prodelta-slope with incised submarine canyons, interfan shelves, and fan-deltas with possible alluvial equivalents. Bathonian to Oxfordian strata are subdivided into corresponding broad facies assemblages which include shelf, fan-delta, prodelta-slope, and submarine fan. The depositional model was modified by Ricketts and Evenchick (1991) to include deep basin, coastal, and submarine canyon/gully assemblages (formerly part of the prodelta-slope assemblage). Table 1 presents an updated version of the depositional model.

Many of the facies assemblages described by Ricketts (1990) and Ricketts and Evenchick (1991) were observed and documented within the Todagin Mountain-Tsatia Mountain region (Figure 2). All facies assemblages listed in Table 1, excluding the deep basin facies assemblage, are represented there. The corresponding lithofacies identified within the present study area are also shown in Table 1. The section below contains detailed descriptions and interpretations of these lithofacies. It should be noted that genetic terms are used throughout this chapter to facilitate comparison with the regional depositional model mentioned above. The corresponding non-genetic terminology for each of the identified assemblages and facies is shown in Table 1.

Table 1.

Subdivisions of the depositional model proposed by Ricketts and Evenchick (1991) for the northern Bowser Basin. Facies assemblages and lithofacies observed in this study are outlined in bold italicized lettering. Each assemblage and facies is qualified by a non-genetic lithologic definition. Interpretive (genetic) terms are used throughout the text in order to facilitate comparison with the depositional model developed by Ricketts and Evenchick (1991).

4.2 Ashman Formation

4.2.1 Submarine Fan Assemblage

4.2.1.1 Upper (Inner) Submarine Fan Facies

Texturally distinct conglomerates are assigned to a unit that occurs along the northern edge of the study area. This unit pinches and swells laterally for approximately 2 km, ranging in thickness from 20 m where it crosses section EPG-I (29-49 m; Appendix I) up to approximately 40 m just east of the section. The unit is bounded above and below by dark grey- and buff-weathered units of the slope facies (discussed below). Conglomerate beds range in thickness from 25 cm up to 2 m and are separated by 5-10 cm planar beds of siltstone to fine-grained sandstone. Bed contacts are sharp and exhibit planar to undulatory bases and tops. Internally, conglomerate beds are massive to inversely graded (Plate 3a) and sorting is very poor to moderate; clast imbrication is weakly developed in the inversely graded beds. Clast- and matrix-supported varieties exist in equal proportions, with framework-to-matrix ratios ranging from 4:1 to 1:1. Framework and matrix components are composed of various proportions of chert (up to 80%), buff- and grey-weathered mudstone to siltstone rip-up clasts (up to 25%), fragmented pelecypods and gastropods (up to 5%), coalified wood fragments (5%), and minor felsic volcanic clasts (< 5%). In moderately sorted beds, the chert-rich matrix ranges from predominantly fine-grained to medium-grained sandstone, whereas mud- and silt-sized matrix is common in massive, poorly sorted beds. The abundance and size of some rip-up clasts is quite pronounced.

Plate 3.

- A. Close-up of inverse grading in a debris flow unit, approximately 100 m east of section EPG-I at 29-49 m. Note the infilled basal scour at the bottom of the photograph. This unit contains two inversely graded beds.
- B. Oblique view to the southeast of a large, rounded mudstone to siltstone ? block (black arrow) in the upper submarine fan facies of the submarine fan assemblage.
- C. Close-up of flame structures in very fine-grained sandstones and siltstones of the slope facies of the prodelta-slope assemblage (section EPG-I at 150 m). Flame structures verge consistently to the south.

They occur up to boulder size (Plate 3b) and range in shape from subangular spheres to elongate laths displaying faint internal bedding. These large clasts predominantly "float" in the upper parts of the stratigraphically higher conglomerate units of the member and show no preferred orientation.

4.2.1.1.1 Interpretation

The overall geometry of the conglomerate member as well as the chaotic and disrupted internal fabric suggests that sediment-gravity processes played a major role in the formation of the member. Specifically, the relatively high mudstone matrix content and the association of marine fossils of probable shelf origin, chaotic internal disorganization, and localized inversely graded, weakly imbricated beds suggest that subaqueous debris flows, as defined by Middleton and Hampton (1973, 1976), were the main depositional mechanisms. Inverse-grading in some beds suggests that cohesionless debris flows or grain flows may also have played a role in the depositional process (cf. Lowe, 1979, 1982). Additional evidence of grainflow processes include the presence of weakly developed a-axis parallel-to-flow imbrication within inversely graded beds (Nardin *et al.*, 1979).

The morphology and internal characteristics of the conglomerate member described above closely resembles submarine fan Facies A described by Mutti and Ricci Lucchi (1972) and Ricci Lucchi (1975). Facies A deposits are mainly formed by subaqueous debris flows and are commonly located in the upper

submarine fan environment (Mutti and Ricci Lucchi, 1972; Ricci Lucchi, 1975; Howell and Normark, 1982; Shanmugam and Moiola, 1988). They are also common within channel systems on continental slopes which feed into submarine fans (Shanmugam and Moiola, 1988). The presence of outsized boulders is misleading, and does not necessarily indicate that the deposit is proximal to the source. Subaqueous debris flows can travel great distances on low-angle slopes, and depending on their thickness, can transport large masses (Nardin *et al.*, 1979). Therefore it is difficult to speculate whether this deposit reflects the base-of-slope or whether it was deposited higher up on the slope within a feeder channel. However, the lack of wave-generated structures suggests that deposition occurred below storm wave base.

4.2.2 Prodelta-Slope Assemblage

4.2.2.1 Slope Facies

Dark grey siltstones monotonously interbedded with buff-weathered mudstones and very fine-grained sandstones were observed in all stratigraphic sections except EPG-IV (EPG-I, 0-1052 m; EPG-II, 0-432 m; EPG-III, 0-450 m; EPG-V, 0-356 m; and EPG-VI, 0-202 m). Its striped appearance (Plate 2a) is typical of parts of the Ashman Formation, originally defined by Tipper and Richards (1976). Most beds are laterally continuous and average 2-5 mm in thickness, but beds up to 20 cm occur rarely. Normal grading is common, with tan weathered fine-grained sandstones grading to grey siltstones and shales. Each shale bed is abruptly overlain by the sandstone base of the next bed. Pinch and

swell of individual beds is common; in extreme cases, beds appear boudinaged. Sporadic flame structures verge consistently to the south (Plate 3c).

Microscopically, most textures within the siltstones and very fine-grained sandstones (chert-lithic arenites) are obscured by a pervasive calcite cement. Sorting of grains tends to be poor to moderate and the textural maturity is immature to submature. Most of the siltstones display framework-to-matrix ratios of 50:50, based on a cutoff size of less than 20 microns for the classification of matrix. The dominant components of the siltstones and very fine-grained sandstones are subround to angular massive chert (20-65%), monocrystalline quartz (5-10%), feldspar (up to 5%), opaques (1-3%), glauconite (1-2%; Plate 4a,b), and calcite cement (up to 40%). Minor amounts of authigenic chlorite (5-7%) rim the surfaces of framework components. Thin section and microprobe analyses indicate that the 0.5-1.0 cm thick buff to beige weathering rinds observed in approximately 10% of the beds are rich in amorphous iron oxides.

Extensive bioturbation occurs sporadically throughout units of the slope facies. In many units, primary bedding surfaces which are defined by concentrations of iron oxide staining are obscured or completely destroyed by bioturbation. Due to the effects of carbonate cement, only the outlines of individual traces are preserved (Plate 4c).

Plate 4.

- A,B. Photomicrographs of glauconite in very fine-grained sandstones of the slope facies of the prodelta-slope assemblage.
- C. Photomicrograph of trace fossils (filter feeders?) within siltstone of the slope facies of the prodelta-slope assemblage.
- D. Ripple cross-stratification within fine-grained sandstone of the overbank facies of the prodelta-slope assemblage (section EPG-V at 76 m). Foresets dip toward the south/southwest (see Figure 8).

4.2.2.2 Overbank Facies

Intermittent sandstone units within the slope facies, ranging from 2-22 m thick, are distinguished from the relatively fine-grained rocks by positive relief of the sandstone units (Plate 2a). The tan to light grey very fine- to fine-grained sandstone units are laterally continuous (up to 50 m), averaging 5-30 cm thick. Individual beds display parallel to wavy bedding as well as pinch and swell geometries; ripple cross-stratification is common (Plate 4d). Rare siltstone rip-up clasts up to 3 cm long are elongate, rounded, and oriented parallel to bedding.

4.2.2.3 Interpretation of the Slope and Overbank Facies

The occurrence of slope and overbank facies consistently below the shelf assemblage indicates that these deposits formed on the slope outboard of the shelf environment. Lithologically similar prodelta deposits laterally associated (outboard) with fan- and braid-delta accumulations were not observed within the study area, although such deposits were observed by Ricketts (1990) in other parts of the basin.

The fine-grained nature of the slope facies suggests low sedimentation rates. According to Ricketts (1990), most of the sediments were introduced to the deeper, quieter portions of the basin via submarine fan systems. On the slopes, inter-submarine channel regions would receive most sediments from hemi-pelagic and pelagic sources. Turbiditic beds incorporating very fine-grained sandstones and siltstones may represent overbank deposition from submarine channelways.

Rip-ups and ripple cross-stratification in the overbank lithofacies may represent overbank deposits where sediment supply rates were increased. Overbank deposits probably accumulated as minor lobes or sheets where slope topography permitted. Vertical and lateral repetition of the overbank facies reflects a dynamic depositional system with frequent changes in sedimentation rates and lateral migration of submarine fan channelways. Consistent southward vergence of flame structures and slump folds suggests a southerly dipping paleoslope (Green, 1991).

4.2.2.4 Slump Facies

Numerous soft-sediment deformation structures exist within the slope facies throughout the northern part of the Bowser Basin. Slides and slumps are the most prominent features on the slope in addition to small-scale deformational structures such as flame structures, convolute bedding, synsedimentary faults, sandstone dykes, and boudinaged beds.

According to Ineson (1985), a slide is a coherent mass of sediment which is transported downslope along a basal detachment (decollement). Slumps, the more highly deformed equivalents of slides, are characterized by concave-up detachments and display evidence of rotational failure (Woodcock, 1979; Postma, 1984a). In the study area, slides range from low-angle stratigraphic discordances to abruptly truncated, coherent sandstone and conglomerate units up to 20 m thick and 250 m in length (Plate 5a). Most slumps in the slope facies are

Plate 5.

- A. Oblique view to the north/northeast of a large slide structure within the slump facies of the prodelta-slope assemblage, approximately 1 km west of section EPG-IV.**
- B. Close-up of convolute bedding within the slope facies of the prodelta-slope assemblage (section EPG-II at 232.5 m). Convolutions verge to the south.**

composed of relatively coarse units (fine-grained sandstone to conglomerate) within shaly host rocks. Slide contacts are sharp and truncate adjacent bedding planes. Synsedimentary faults (most penetrating less than 5 cm into the sediment), flame structures, and convolute bedding occur in sediments directly underlying slump structures (Plate 5b). The large slide structure shown in Plate 5a as well as flame structures distributed throughout the slope facies consistently verge to the south. Kaip (1991) has provided a detailed account of synsedimentary deformation structures within the prodelta-slope assemblage, including criteria for distinguishing them.

4.2.2.4.1 Interpretation

The presence of both small- and large-scale soft-sedimentary deformation structures throughout the slope facies suggests an unstable depositional environment. Slumps and slides are believed to occur when the shear stress produced by gravity becomes greater than the frictional shear stress of the sediment along a potential failure plane (Lewis, 1971). Three main causes of slumping and sliding include seismic activity, increased pore fluid pressures within sediments, and slope oversteepening (Davis, 1983). Seismic activity is associated with tectonically active zones while increases in pore fluid pressures are commonly attributed to increased sedimentation rates (Helwig, 1970). Finally, oversteepening of slopes can result from increased sedimentation rates, undercutting of slopes by current erosion, and slope sediment thickening by slides

and slumps originating farther upslope (Lewis, 1971).

The abundance of load structures suggests at least periodic, rapid sedimentation rates within the slope environment. Large-scale slides and slumps are interpreted to be the causative agent for soft-sedimentary structures underlying them. The theory of Eisbacher (1985), that the Bowser Basin initially developed as a foredeep in response to overthrusting of Cache Creek onto Stikinia, implies a tectonically active environment. Seismic activity would be relatively frequent in this setting (Kaip, 1991). Southerly vergence of both large- and small-scale deformation structures within the slope facies suggest a south-dipping paleoslope, but the angle of repose of the slope is difficult to establish. Although some have reported that slumps and slides occur in environments with relatively steep slopes (e.g. Davis, 1983), others have claimed that slumping can occur on slopes as gentle as 1-5° (Rupke, 1986; Lewis, 1971). Kaip (1991) suggested that the paleoslope angle ranged from less than 1° in the regions characterized by the slope facies, to greater than 5° near the shelf/slope break and submarine canyon/gully assemblage.

4.2.3 Submarine Canyon/gully Assemblage

4.2.3.1 Channel Facies

Significant amounts of conglomerate are spatially associated with the slope facies. On Tsatia and Todagin mountains, conglomerate bodies occur as wedges from 5 m thick and 20 m wide (cross-section viewing down depositional axis) to coalesced offstacked wedges 100 m thick and 400 m wide. Conglomerate bases

are extremely abrupt and typically occupy scours cut into the underlying fine-grained rocks (Plate 6a). Conglomerate bodies are rarely associated with coarsening-up trends within the sedimentary sequence, and channel margins tend to pinch out laterally. Thick conglomerates (> 30 m) represent a series of offstacked wedges overlapping one another. In a few places, they are separated by thin units of siltstone and fine-grained sandstone, generally less than 50 cm thick. Distinct internal stratification in the conglomerates is manifested by alternating discontinuous fine- to medium-grained sandstone beds (Plate 6b). Beds range from 5 to 40 cm thick. Bedform geometries are typically planar to slightly wavy, and rarely contain low-angle foresets; discontinuous sandstone beds commonly occur between foresets. Trough cross-stratification is absent. Inverse to normal grading is common in individual conglomerate beds, although ungraded varieties exist. Clasts are moderately to well-sorted; subrounded to well-rounded, tabular clasts define a pronounced a/b plane imbrication. Most imbricated clasts display a-axis orientation parallel to flow direction and dip upcurrent (toward the north-northeast). Clast frameworks are intact to condensed; framework to matrix ratios range from 4:1 to 3:2. The matrix is composed of fine- to medium-grained, poorly sorted chert-rich sandstone.

Lateral exposure of a 2 to 50 m thick conglomerate unit, herein informally named the "Waterfall Conglomerate", in the upper part of the slope facies on Tsatia Mountain (Plate 7)(EPG-V, 356-363 m; EPG-VI, 202-214 m), reveals that the character of the channel conglomerates changes down depositional axis

Plate 6.

- A. View to the southeast of channel conglomerates of the submarine canyon/gully assemblage. Note the relief on the abrupt basal contact (section EPG-I at 484 m).**
- B. Close-up showing well-stratified conglomerate and sandstone beds. Note the clast imbrication with clasts dipping to the right (section EPG-I at 900 m).**

Plate 7.

Panoramic mosaic from southeast to northeast of the west side of Tsatia Mountain. The black arrows point to the Waterfall Conglomerate which extends laterally for 2 km in this plate.

(south-southwest). At the north end of Tsatia Mountain, the conglomerate units are poorly sorted, indistinctly bedded, and are pervasively rusty weathered (Plate 8a). To the south, the conglomerates are weathered grey, display clast sorting and are more distinctly stratified (Plate 8b,c). At the south tip of Tsatia Mountain, approximately 2 km along the unit, clasts within conglomerate beds are strongly imbricated (a-axis parallel to flow) and dip upstream, clast sizes are relatively small, and the proportion of interstratified sandstone increases. Perpendicular to the depositional axis, the conglomerates show a change in geometry from sheetlike to a series of lobes.

In thin section, the conglomerates are submature and typically display moderate sorting. Packing of the framework clasts, which range in size from less than 1 cm to greater than 10 cm (mean clast size: 4-6 cm), is intact to dispersed. Grain-boundary contacts between intact framework grains are typically straight, tangential and less commonly, sutured. The framework and matrix are composed of relatively similar proportions of massive chert (up to 40%; Plate 9a), radiolarian-bearing chert (up to 30%, Plate 9b), and rock fragments composed of monocrystalline feldspar (up to 5%) and very fine-grained silty chert or altered volcanics (up to 20%; Plate 9c). Cpaques consisting of iron oxide? and pyrite (Plate 9d) make up to 3% of the rock. Approximately 1-2% of the conglomerates and sandstones are composed of detrital glauconite (Plate 4a,b) that displays a fibrous habit and in most places has been deformed between other framework grains.

Plate 8.

- A.** View to the north of the indistinctly bedded, poorly sorted Waterfall Conglomerate, at the northwest side of Tsaia Mountain (section EPG-VI at 204 m).
- B.** View to the northwest of the Waterfall Conglomerate, approximately 1 km down depositional axis from Plate 8A. Note the distinct stratification within the conglomerate unit. The white part of the Jacob staff, at the bottom of the plate, represents 0.5 m.
- C.** Close-up of Plate 8B. Lens cap (right centre) for scale.

Plate 9.

- A. Massive chert clast within a chert-lithic conglomerate. Internal fractures within the clast are filled with sutured quartz cement.**
- B. Radiolarian-bearing chert clast with stylolites.**
- C. Andesitic (?) volcanic clast with feldspar phenocrysts in a fine-grained groundmass.**
- D. Fracture within a chert clast containing polygonal quartz and opaque iron oxide cement.**
- E. Micro-quartz and chalcedony cement.**
- F. Interclast pore rimmed by micro-quartz, mega-quartz and laumontite (in order from the clast rim to the centre of the pore).**
- G. Chlorite cement (c) infilling an interclast pore. Rimming cement is quartz (q).**
- H. Monocrystalline feldspar clast showing chlorite alteration along cleavage traces.**

Quartz, chlorite and iron oxide cements occur in various proportions throughout the conglomerates and sandstones. The predominant cement is silica, which occurs as chalcedony (Plate 9e), and mega- as well as micro-quartz (Plate 9f). It occurs along grain boundaries, as infillings within voids, and along fractures within clasts (Plate 9a). With the aid of microprobe analysis, the second most abundant cement was identified as chlorite (Kaip, 1991). Additional electron microprobe analyses indicated that this chlorite group mineral is penninite, according to the nomenclature of Hey (1954). It occurs as patchy masses within all clast types, especially within the fine-grained volcanic clasts. Chlorite aggregates also commonly form within intergrain pores (Plate 9g), and tend to follow contacts between polygonal quartz cement. Detrital monocrystalline feldspars are particularly susceptible to chlorite and sericite alteration along twin boundaries and cleavage traces (Plate 9h). The third cement is composed of an opaque iron oxide. The first two cements may have formed penecontemporaneously because they cross-cut each other in different thin sections. No pattern is evident in the relative timing of cements in relationship to stratigraphic level. Laumontite has been documented as a fourth cement in a conglomerate in section EPG-III, at 1296-1305 m. This zeolite mineral is a late pore-filling cement that has replaced quartz (Plate 9f).

4.2.3.1.1 Interpretation

Channel shaped conglomerates of the submarine canyon/gully assemblage represent infillings of incised canyons and gullies (Ricketts, 1990). Coarse clastics shed from the north are inferred to have bypassed the shelf accumulations and funnelled down the slope along depressions which may represent slump scars on the slope (Ricketts, 1990; Green, 1991). In a fluidized environment, coarse coastal sediments can be expected to eventually become well sorted and stratified with increased time and distance travelled down canyons; this is the case with the Waterfall Conglomerate.

Travelling down the gullies as sediment-gravity flows, most of the coarse-grained sediments likely originated from slumps on the unstable, overloaded fan-delta front. The change in fabric of the Waterfall Conglomerate from relatively massive, poorly sorted, and ungraded, to well-graded and stratified, is similar to the downslope transition of fan-delta slope deposits described by Postma (1984b). The process referred to by Postma (1984b) as gravity-winnowing may be the mechanism responsible for selective segregation of grains from unstable sediment masses such as sediment-gravity flows. Initially, the sediment-gravity flows behave plastically. With time and distance travelled down the slope, however, the flows transform into high-density turbulent flows. It is during this stage of gravity-winnowing that imbrication, showing a-axis parallel to flow, can be developed (Postma, 1984b). This explains the predominance of a-axis imbrication in the channel conglomerates, analogous to resedimented conglomerates showing similar

imbrication patterns (cf. Walker, 1984). If the clasts were transported in a fluvial environment, imbrication would be characterized by a-axis transverse to flow due to bed rolling during transport (Hein, 1984; Nemec and Steel, 1984; Walker, 1984); clearly, this was not the case.

Channel conglomerates of the submarine canyon/gully assemblage tend to occur as coalesced units on Tsatia and Todagin mountains; this may reflect significant provenance from the north. The lateral offstacking of individual wedges probably reflects the sinuosity of such routes, and the repetitive vertical stacking probably records periodic increases in rates of sediment influx. However, the latter can also be due to changes in sea-level related to eustasy or changes in the rate of basin subsidence.

4.2.4 Maximum Clast Size (MCS) versus Bed Thickness (BTh)

Sediment-gravity flows are inferred to have been responsible for deposition of the channel conglomerates of the submarine canyon/gully assemblage. These types of flows can range from fully turbulent turbidity currents (cf. Walker, 1984) to debris flows (cf. Lowe, 1982). The latter are subdivided into cohesive debris flows and cohesionless debris flows or grainflows (Nemec and Steel, 1984). Cohesive debris flows tend to behave plastically and exhibit relatively high matrix strength, whereas grainflows maintain clast support by frictional effects such as dispersive pressure (Nemec and Steel, 1984).

One of the empirical means to distinguish the type of sediment-gravity flow is the positive linear correlation that exists between bed thickness (Bth) and maximum clast size (MCS), which represents the largest clast that can be held in suspension during the life of the flow (Nemec and Steel, 1984). This type of correlation is not observed in fluvial environments where different depositional processes occur (Nemec and Steel, 1984). A comprehensive account of the MCS versus BTh technique has been reviewed by Nemec and Steel (1984); a description of the methodology used for the present study is in Appendix II.

The correlation coefficient (R, Figure 6) for the MCS versus BTh plot is lower than those derived from similar studies of conglomerates (cf. Koster and Steel, 1984). This is likely due to the fact that measurements were taken from different locations throughout the conglomerates of the channel facies. As previously mentioned, the internal fabric of the Waterfall Conglomerate changed downslope in response to an evolving sediment gravity flow through the process of gravity winnowing. This reflects a spectrum of gravity flow mechanisms such as cohesive flows in proximal channel areas changing to grainflows in more distal regions. A lower than expected correlation coefficient occurs because there is a mixture of different fabric types that represent different flow mechanisms. This is unlike other studies where fabric subdivisions were identified to delineate sediment-gravity flow types (Nemec and Steel, 1984; Mustard, 1990). Although the results are not as conclusive as those cited above, they still suggest that a gravity-flow mechanism was dominant.

Figure 6.

Plot showing the relationship between maximum clast size (MCS) and bed thickness (BTh) for channel conglomerates of the submarine canyon/gully assemblage. R = Pearsonian coefficient value, n = sample number, line represents the best-fit linear regression for the data.

4.3 Undivided Bowser Lake Group/Sandy Lithosome

4.3.1 Shelf Assemblage

The shelf assemblage, informally known as the sandy lithosome (Ricketts and Evenchick, 1991), rests conformably over strata of the Ashman Formation. On Todagin Mountain, the shelf assemblage was only observed at section EPG-II, 432-552 m. All four stratigraphic sections on Tsatia Mountain contain strata representative of the shelf assemblage (EPG-III, 460-1512 m, 1780-1810 m; EPG-IV, 0-116 m, 207-384 m; EPG-V, 356-1004 m, 1010-1080 m; EPG-VI, 202-860 m). In section EPG-III, over 1000 m of shelf deposits interfinger with rusty weathered conglomerates of the fan delta and coastal assemblages (discussed in next section). Section EPG-IV contains the thinnest shelf assemblage, measuring 116 m; this is followed by 200 m of strata in EPG-II. Sections EPG-V and VI exhibit similar shelf thicknesses of approximately 640 m.

The shelf assemblage contains cyclical coarsening-upward packages of siltstone, micritic sandstone (nomenclature after Mount, 1985), sandstone (chert-lithic arenites), and sporadic caps of chert-lithic conglomerate. Individual cycles range from 5 m up to 30 m thick (Plate 10a). Cycle bases are characterized by recessive thinly bedded shales containing minor amounts (< 10%) of mudstone and chert pebbles. Thinly layered micritic sandstone, characterized by alternating laminations of micrite and silt to fine sand-sized siliciclastic grains, sporadically occur throughout lower parts of cycles. Grain size and bed thickness increase upward. Toward the tops of cycles, fine- to medium-grained sandstone contains

Plate 10.

- A. View to the east of the west side of Tsatia Mountain (ridge in the foreground, leading up to the peak, represents section EPG-VI). The striped appearance is due to the cyclical nature of the shelf assemblage.
- B. Large concretions in shelf assemblage sandstones. The Jacob staff is 1.5 m long.
- C. Low-angle planar cross-stratification in rusty weathered medium-grained sandstones of the shelf assemblage (section EPG-V at 1104 m).

abundant concretions that range from perfectly spherical to ovoid. Ranging from 1 cm to 1 m in diameter, they are post-compactional and tend to obscure primary sedimentary structures (Plate 10b). An internal examination of several concretions shows that organic material such as plant, wood, and fossil fragments acted as the nuclei for concretion growth. Low-angle planar cross-stratification is a common bedform in coarse-grained parts of cycles (Plate 10c); wavy cross-stratification, planar beds, and hummocky cross-stratification (HCS) are also common; because of the geometry of HCS, only isolated parts could be identified at any given locality. Asymmetric ripples (Plate 11a) and siltstone rip-up clasts (Plate 11b) in fine- to medium-grained sandstones occur in the middle and upper parts of coarsening-upward cycles. Fossil-rich beds near the tops of cycles are up to 20 cm thick, and contain up to 60% disarticulated pelecypods, belemnites, ammonites and rare gastropods (Plate 11c). Sporadic rounded chert pebbles and sandstone rip-ups are associated with these beds. Apart from the cycle-capping conglomerates described by Ricketts (1990), poorly sorted chert-lithic pebble conglomerates occur sporadically throughout the shelf assemblage. These 15-25 cm thick conglomerate beds occur in shale and sandstone components of cycles. Characterized by abrupt, undulatory erosive bases with pebble-filled scours and flat tops, the conglomerates commonly grade into sandstones. Frameworks are intact to condensed, and clasts show extreme ranges in size, shape and composition (Plate 11d). The largest clasts tend to 'float' in the middle of beds. Rounded siltstone intraclasts and fossil fragments (up to 15% each) are locally

Plate 11.

- A. Asymmetric ripple cross-lamination in fine-grained sandstones of the shelf assemblage (section EPG-VI at 396 m).**
- B. Siltstone to fine-grained sandstone intraclasts in a calcareous sandstone unit (section EPG-VI at 440 m).**
- C. Fossil-rich bed within fine- to medium-grained sandstones of the shelf assemblage (section EPG-VI at 716.5 m).**
- D. Poorly sorted chert-lithic pebble conglomerate bed within sandstones of the shelf assemblage (section EPG-VI at 384.8 m).**
- E. Micro-stylolites within a massive chert clast.**
- F. Multiple fracture patterns within a radiolarian-bearing chert clast suggests a complex deformational history.**
- G. Calcite cement surrounding massive chert clasts within a coarse-grained chert-lithic sandstone.**

abundant. The matrix is composed of siltstone and medium-grained, poorly sorted chert-rich sandstone. Matrix content tends to increase toward the tops of beds. Framework-to-matrix ratios are typically 70:30.

Toward the top of each shelf assemblage section, the abundance of carbonaceous plant fragments shows a subtle increase. Plant fragments, which rarely compose more than 3% of the rocks, are generally less than 1 cm thick, and occur throughout the siltstone and sandstone beds. The ratio of rusty to grey weathering in sandstones also increases toward the conglomerates of the fan delta/coastal assemblages. This is particularly obvious in sandstones at the tops of cycles.

In thin section, shelf assemblage lithologies display compositions similar to those described in channel conglomerates of the submarine canyon/gully assemblage. However, some notable differences in composition include the introduction of chlorite rock fragments as well as the presence of micritic interbeds within siltstones and fine-grained sandstones. Typically, the sandstones show framework-to-matrix ratios of 90:10 and higher, whereas mixed carbonate and siliciclastic rocks containing up to 35% micrite display smaller ratios averaging 60:40. The framework grains are intact, and grain contacts are predominantly point, straight, tangential, and rarely sutured. Grain shapes range from rounded to angular with most being sub-angular. The chert-lithic arenites are typically composed of up to 40% chert (because of limited grain size, massive and radiolarian-bearing chert could not be distinguished), 30% chlorite rock

fragments, 15% altered volcanics or silty chert, up to 15% strongly iron-oxide stained rock fragments (composition indistinguishable), 10% monocrystalline quartz, and 5% detrital feldspar. Many chert grains contain fractures which are now filled by unstained quartz cement; fracturing and the infilling quartz cement predates final deposition (Plate 9a). Many of the chert grains display compaction features such as microstylolites (Plates 9b, 11e) and multiple fracture patterns (Plate 11f).

Calcite, silica, chlorite and to a lesser degree, iron oxide, constitute the post-depositional cements. In most cases, quartz is the initial cement. However, calcite cement (Plate 11g) which can constitute as much as 40% of the rocks, tends to obscure the exact timing of the other cement components. This cement is sporadic and is commonly confined to individual units and beds.

4.3.2 Interpretation of Shelf Assemblage

Strata of the prodelta-slope and submarine canyon/gully assemblages are conformably and gradationally overlain by deposits which are interpreted to have formed in a marine shelf environment (Ricketts, 1990; Green, 1991). The conformable transition from the underlying deposits is abrupt, and may represent the shelf-slope break (Ricketts, 1990). The lateral variation in shelf thickness over a relatively small region indicates either a complex margin with different shelf widths, or regions of selective preservation (e.g. inter-fan shelves; Green, 1991). One explanation for variation in the shelf geometry involves delta

progradation. The part of the shelf receiving the brunt of sediment from a delta would increase in thickness and in surface area (Bouma *et al.*, 1982). The adjacent parts of the shelf would be marginally affected; the result is a shelf of variable thickness and width.

Several lines of evidence including facies associations, sedimentary structures, large-scale sedimentological trends, macrofossils, and mineral compositions collectively suggest a shallow-marine shelf environment. For example, strata listed in the preceding descriptive section are underlain by slope and submarine canyon/gully deposits (deposited in relatively deep parts of the basin). Fan-delta and coastal deposits overlie, and are adjacent to, the strata in question. By facies association, it therefore is reasonable to suggest that strata bounded by the above environments would be deposited in a shelf setting. The cyclic coarsening-upward trend and overall shoaling of this assemblage (Eisbacher, 1981) is typical of a shallow marine shelf which was sensitive to repeated transgressive and regressive events (cf. Cant and Hein, 1986). The upward increase in carbonaceous matter in concert with moderate- to well-sorted conglomerates toward the tops of some cycles may indicate proximity to coastal regions (Ricketts, 1990; Green, 1991). Moreover, the appearance of glauconite in some of the lithologies suggests a marine environment (Johnson and Baldwin, 1986). Sedimentary structures such as HCS, asymmetric ripples and rip-up clasts indicate increased energy levels in parts of the shelf environments (discussed below). Finally, the presence of pelecypods and ammonites indicate shallow-

marine conditions (T. Poulton, pers. comm., 1992).

The shelf environment can be broken down into several facies environments: nearshore, inner shelf and outer shelf. These facies are subdivided on the basis of sedimentary structures, textural changes such as grain size, macrofossil occurrences, and ichnofacies (Johnson and Baldwin, 1986). The transition from the nearshore to the outer shelf is generally characterized by a reduction in grain size, an increase in bioturbation, and a decrease in sediment sorting (Johnson and Baldwin, 1986). Important features of the shelf environment are the fairweather and storm wave bases, which, according to Walker (1984), range from 5 m to 15 m. Above this depth, sediments are continually being reworked by wave action. Below the fairweather wave base, conditions on the shelf are normally quiet, but during storms, the effects of waves on the shelf floor extend much deeper; this lower limit is the storm- or foul-weather-wave base (Walker, 1984). Dott and Bourgeois (1982) have proposed that HCS can form in water depths as great as 80 m; this extends the lower limit of the storm-wave base to at least this depth.

4.3.2.1 Outer Shelf Facies

The lower parts of the coarsening-upward cycles within the shelf assemblage are predominantly composed of siltstone, very fine-grained sandstone, and sporadic layers of micritic sandstone. The lack of evidence of fairweather reworking and the high fractions of siltstone suggest deposition below fairweather

 

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wave base (Atkinson et al., 1986; Johnson and Baldwin, 1986; Decelles, 1987).

Micritic sandstone may indicate conditions which favoured carbonate production separated by influxes of siliciclastic detritus. This process of punctuated mixing can occur when storms periodically distribute siliciclastic detritus into regions of carbonate production (Mount, 1984), regions that may represent an outer shelf or quieter parts of an inner shelf facies.

4.3.2.2 Inner Shelf Facies

A relative increase in the proportion of sandstone toward the middle of coarsening-upward cycles suggests higher energy levels. The occurrence of asymmetric ripples in several localities (EPG-IV, 248 m; EPG-V, 495 m) in association with a relative increase in macrofossils (particularly pelecypods) and the appearance of carbonaceous matter, collectively suggest inner shelf conditions. The common occurrence of bedforms resembling parts of HCS also indicates deposition above storm wave base (see below). The presence of siltstone rip-up clasts in fine- to medium-grained sandstones indicates reworking of partially consolidated sediment. This suggests episodic high energy events (storms) separated by intervals long enough to initiate consolidation; again, this suggests deposition below fairweather but above the storm wave base.

4.3.2.3 Nearshore Facies (shoreface ?)

The upper parts of coarsening-upward cycles (EPG-V, 789-801 m; EPG-VI, 420-432 m, 774-787 m) are characterized by massive and low-angle planar cross-stratified fine- to medium-grained sandstone. These sedimentary structures in association with increased carbonaceous matter (indicating proximity to the coast), and moderately- to well-sorted pebble conglomerates (high energy) are indicative of nearshore to shoreface environments (Reinson, 1984). Pervasive rusty weathering typical of the upper parts of the stratigraphically higher coarsening-upward cycles may indicate oxidizing conditions (to be discussed). This facies is transitional and interdigitates with the rusty lithosome described in the next section.

4.3.3 Effects of Storms on Shelf Environments

The effects of storms strongly influence the sedimentary characteristics of shelf environments. This is particularly significant since 80% of the world's modern shelves are storm-dominated (Swift *et al.*, 1986). During storms, significant amounts of sediment are transported and reworked by wave-induced storm surge (Richards, 1986) and offshore currents (i.e. ebb-flow; Aigner, 1985; Swift *et al.*, 1986). In storms of moderate strength, wind-driven currents appear to be the main sediment transporting mechanism. However, ebb-flow currents appear to be the dominant process in offshore sand sheet progradation during major storm events (Nelson, 1982). It appears that storms influence all parts of

the shelf environment including the outer shelf.

The sporadic poorly sorted chert-lithic pebble conglomerates and siltstone rip-up beds described in the preceding section are similar in character to shelf storm deposits described by Lewis (1980) and Bouma *et al.* (1982). The shell hash layers are also interpreted as representative of storm events (Bouma *et al.*, 1982).

The identification of HCS in numerous localities within the shelf assemblage suggests deposition above the storm wave base (Walker, 1984). Most researchers agree that HCS is associated with storm-influenced shelf deposits (Harms *et al.*, 1975; Walker, 1979; Brenchley *et al.*, 1986; Decelles, 1987). Research in modern environments indicates that HCS is formed by the combined effects of oscillatory flow (wave-generated) and superimposed unidirectional flow; these conditions are commonly produced during storms (Swift *et al.*, 1983; Greenwood and Sherman, 1986; Decelles, 1987).

4.4 Undivided Bowser Lake Group/Rusty Lithosome

4.4.1 Fan-Delta and Coastal Assemblages

Rusty weathered chert pebble- to cobble-conglomerates comprise several lithofacies within the Bowser Lake Group. They are intimately associated with, and are laterally equivalent to, the shelf assemblage already discussed.

Approximately 64 m of the rusty lithosome are represented in section EPG-II (552-616 m) on Todagin Mountain; rusty conglomerates and sandstones also occur in all four measured sections on Tsatia Mountain (EPG-III, 1014-1304 m, 1512-

1780 m; EPG-IV, 116-207 m; EPG-V, 1004-1010 m, 1080-1214.5 m; EPG-VI, 202-216 m, 495-509 m, 860-1042.5 m). These conglomerates are sheetlike and extend for several kilometres within the Tsatia Mountain region (Plate 12a); amalgamated conglomerate units on the north side of Tsatia diverge southward into the shelf assemblage. A 14° discordance between the rusty braid-delta facies and the fan-delta slope facies can be seen 1 km southwest of section EPG-IV (Plate 12b). Rusty weathered conglomerate units (some up to 40 m thick) are separated by rusty weathered coarsening-upward sequences of fine- to medium-grained sandstone which commonly contain HCS and low-angle planar cross-stratification (shoreface facies). Toward the top of Tsatia Mountain, a unit of carbonaceous shales less than 1 m thick occur between the upper two units of rusty weathered conglomerate.

Diverse internal fabrics of the rusty conglomerates differ from fabrics of the channel conglomerates of the submarine canyon/gully assemblage. Bed thickness ranges from 10 to 40 cm. Bedforms include planar beds, tabular-planar cross-stratification, and trough cross-stratification inclined 10-20° (Plate 13); the latter commonly display curved erosion surfaces. Lenticular, concave-upward basal bedform geometries are common. Non-graded beds are abundant, although normal grading exists up foreset dip in some cases. Clast sorting and imbrication are moderate to poor, and outsized clasts are common (Plate 14a). Clasts range from round to subangular and display a wide range of sphericities. Framework-to-matrix ratios are similar to the channel conglomerate facies, but one example

Plate 12.

- A.** Panoramic mosaic from south to southwest showing the sheet-like geometry of rusty weathered conglomerates on Tsatia Mountain. The field-of-view of this photograph is approximately 3 km.
- B.** View looking southeast showing a local discordance between rusty weathered conglomerates (braid-delta facies) and the underlying fan-delta slope facies of the fan-delta assemblage (1 km southwest of section EPG-IV).

Plate 13.

Trough cross-stratification in rusty weathered chert-lithic pebble conglomerates of the braid-delta facies of the fan-delta assemblage (section EPG-III at 1668 m).

Plate 14.

- A. Close-up of an outsized clast in rusty weathered conglomerates of the braid-delta facies (approximately 50 m above section EPG-II on Todagin Mountain).**
- B. Close-up of an openwork conglomerate of the braid-delta facies (same location as plate 14A).**

of an openwork conglomerate was observed in the upper part of a unit above section EPG-II (Plate 14b). Discontinuous sandstone beds occur less commonly than in the channel conglomerate facies, and carbonaceous plant fragments are abundant within the rusty weathered conglomerates. A displaced 2 m x 10 cm tree trunk was observed near the top of section EPG-III. Marine fossils such as pelecypods and ammonites were seen near the peak of Tsatia Mountain during regional mapping in 1989 (Evenchick, pers. comm., 1992); no fossils were observed by the author in the study area.

Microscopically, rusty weathered conglomerates are similar in composition but dissimilar in texture to conglomerates already described in this chapter. Texturally, these conglomerates are more tightly packed than those of the submarine canyon/gully and shelf assemblages. Moreover, a higher proportion of sutured contacts with associated stylolites is documented. Porosity is generally less than 1% except in openwork conglomerates. For example, the openwork conglomerate above section EPG-II display porosities of 15%. As the name suggests, the rusty weathered conglomerates and bounding sandstones commonly display a ubiquitous coating of iron oxides which imparts an intense red coloration to the rocks. X-ray diffraction analysis indicates that the iron staining is composed of hematite and goethite. Quartz and iron oxide make up the primary cements and compose up to 10% of the rocks.

4.4.1.1 Interpretation

Rusty conglomerates and their associations in the Tsatia Mountain area represent a continuation of the overall coarsening-upward and shoaling trend of the Bowser Lake Group in the northern Bowser Basin. They display some of the lithofacies divisions of the rusty lithosome delineated by Ricketts and Evenchick (1991). Discerning the exact transition of the environments within the amalgamated units was in places difficult due to intense hematite staining which tends to mask fabrics within conglomerate units. However, the rusty lithosome appears to represent a spectrum of environments including braided channels of braid deltas, coastal environments such as gravel bars with lagoonal associates, channel-mouth bars of the fan/braid-delta front environment, and possibly deposits of the fan/braid-delta slope.

Extensive discussions of fan- and braid-deltas in the literature have led to numerous attempts to properly define these depositional systems. As a result, a great amount of ambiguity and uncertainty exists as to the proper use of these terms for depositional interpretations. A braid-delta is currently defined as: "all coarse-grained (gravel-rich) deltas that are formed by the progradation of a purely fluvial, braided alluvial-plain system, of whatever origin, into a standing body of water" (McPherson *et al.*, 1988; p. 16). A fan-delta is "a coastal prism of sediments delivered by an alluvial-fan system and deposited, mainly or entirely subaqueously, at the interface between the active fan and a standing body of water" (Nemec and Steel, 1988; p. 7). The basis for distinguishing these two

systems lies in the identification of the appropriate subaerial component; subaqueous and transitional shoreline deposits can be quite similar (McPherson et al., 1988). In this study, the amalgamation of units within the rusty lithosome made the task of distinguishing the subaerial component of the depositional environment difficult. Furthermore, Miall (1977) has pointed out that both lower alluvial fans and proximal braided river systems contain similar braided channel deposits (discussed below); the following interpretations of shoreline and subaqueous deposits are therefore qualified by the term "fan/braid" as the prefix. The term braid-delta is used on the basis of evidence presented in the next section. Ricketts and Evenchick (1991) established the fan-delta assemblage to incorporate the regionally identifiable fan-, braid, and gilbert-delta facies throughout the northern part of the basin.

A MCS versus BTh analysis was carried out within the braid delta facies as a possible means to either prove or disprove the presence of sediment-gravity flows. The correlation coefficient for this study is extremely low (Figure 7), suggesting that sediment-gravity flows did not play a significant role in the depositional process. This is consistent with inference of a braid-delta facies in which fluvial processes were active.

Figure 7.

Plot showing the relationship between maximum clast size (MCS) and bed thickness (BTh) for braid-delta conglomerates of the fan-delta assemblage. R = Pearsonian coefficient value, n = sample number, line represents the best-fit linear regression for the data.

4.4.2 Fan-Delta Assemblage

4.4.2.1 Braid-Delta Facies

The braid-delta facies is contained within the fan-delta assemblage of Ricketts and Evenchick (1991). Several lines of evidence are suggestive of a braid delta environment. Firstly, the sheetlike geometry of the rusty lithosome is typical of braided channel environments (Plate 12a). Fan-delta deposits tend to be laterally discontinuous and wedge-shaped (McPherson *et al.*, 1988). Festooned trough and planar cross-stratified units in section EPG-IV (190-207 m)(Plate 15a) represent the vertical and lateral stacking of gravelly channels and may also represent lateral migration of longitudinal bars within a channel system (Marzo and Anadon, 1988). Similar generalized descriptions of laterally migrating longitudinal bars in braided channel networks include Miall (1978) and Rust (1978). Ito and Masudo (1988) suggested that trough cross-stratification within braided channels represent infillings of inter-bar pools and cross-cutting channels. Moreover, sporadic lenticular sandstone beds may represent accumulations on bar flanks during decreased flow (Ito and Masudo, 1988). Commonly associated with this complex stratification are poorly defined, relatively thin (< 10 m) planar stratified conglomerate beds. These may reflect intermittent sheetflood events on alluvial plains or may represent tops of longitudinal gravel bars (Nemec and Steel, 1984; Rust and Koster, 1984). The presence of outsized clasts may be the result of decreased flow competence in response to increased flow width at the mouths of channels between longitudinal bars (Gloppen and Steel, 1981; Mustard, 1990). Interlayered normally graded and ungraded beds are typical of fluvial systems (Kleinspehn, 1984).

Plate 15.

- A. View to the northeast of stacked planar and trough cross-stratification in rusty weathered chert-lithic pebble conglomerates of the braid-delta facies of the fan-delta assemblage (section EPG-IV at 200 m).**
- B. View to the east of large-scale "foresets" of the fan/braid-delta slope facies of the fan-delta assemblage (section EPG-IV at 172 m).**

The lack of mud- and silt-sized fractions within this part of the rusty lithosome suggests subaqueous reworking and relatively efficient sorting capacity which is typical of braided channel systems (McPherson *et al.*, 1988). These finer fractions would be carried farther along the system and out to the sea.

Openwork conglomerate at one locality indicates rapid accumulation of gravel along foresets. Rapid deposition inhibits matrix infiltration amongst framework components (Rust and Koster, 1984). Sieve deposits are a type of openwork conglomerate which are interpreted to represent matrix-free lobate permeable gravels. The absence of matrix in these conglomerates may be caused by nondeposition, removal by rainwash, or infiltrating water discharge related to deposition (Hooke, 1967). Although they are commonly associated with streamflow processes (Bull, 1972) sieve textures can also form at the front or in levees of debris flows (Hooke, 1967).

One of the primary features which supports a braid delta interpretation is the occurrence of marine fossils within the study area. This clearly suggests that marine deposition was an integral part of the rusty lithosome. This does not preclude the possibility of emergent subaerially exposed regions.

4.4.2.2 Fan/Braid-Delta Front Facies

Fan-delta deposits are composed of three main subaqueous environments : (1) fan-delta front, (2) fan-delta slope, and (3) prodelta (Postma, 1984a). The fan-delta front is the region above wave base and is considered to be the highest

energy zone within the fan-delta system. The fan-delta slope is the region characterized by active resedimentation (mostly sediment gravity flows) on high-gradient slopes. The prodelta is the area beyond the delta slope and consists mainly of silts and muds (Postma, 1984a).

Extensive channel mouth bar deposits can develop in a rapidly prograding system (Marzo and Anadon, 1988). These deposits may become emergent as they prograde into the marine environment toward the fan-delta slope.

Channel mouth bar deposits are intimately associated with shoreface and nearshore deposits. For example, in section EPG-III, 1610 m-1634 m, low-angle planar cross-stratified fine- to medium-grained sandstones (i.e. shoreface) are conformably overlain by planar cross-stratified and subhorizontally stratified chert pebble-conglomerate units. The conglomerate units at this locality are vertically stacked and consist of erosively-based, cross-cutting, lenticular beds sporadically interbedded with laterally discontinuous sandstone wedges. These features are typical of aggrading subaqueous mouth bars (Marzo and Anadon, 1988; Colmenaro et al., 1988; Nemec and Steel, 1984). Furthermore, these geometries may represent small-scale channels and bars within channel mouth bar sequences. These are formed in response to the frictional effluent from fluvially dominant channel mouths (Wright, 1977; Marzo and Anadon, 1988).

4.4.2.3 Fan/Braid-Delta Slope Facies

As indicated above, fan-delta slope deposits consist mainly of sediment-gravity flows (mostly debris flows) on high-gradient slopes. An example of this environment was described by Postma (1984b) who studied the Abrioja Formation of southeastern Spain, a fan-delta sequence containing delta slope deposits. In his study, upper-delta slope deposits were inferred to represent sediment-gravity flows. A unique characteristic of the overall geometry of these deposits is that they are steeply inclined (6-25°) relative to the overlying subaerial deposits. Inclined strata resemble large foresets of Gilbert-style fan-deltas (Postma, 1984b). The large-scale geometry of the bedforms in Plate 15b closely resembles that described above. Furthermore, in Plate 15b, discordant top set beds are identified as those belonging to the braid-delta facies. The giant "foresets" may represent the accumulations of sediment-gravity flows on the fan/braid-delta slope. Conclusive evidence that these deposits represent fan/braid-delta slope deposits is lacking due to masking of internal fabrics by hematite staining. However, some of these bedforms are massive and lack an organized internal fabric. Massive beds are considered to be one of the characteristic bedforms of debris flow deposits (Nemec and Steel, 1984). Therefore, these deposits may represent a fan/braid-delta slope environment.

4.4.3 Coastal Assemblage

4.4.3.1 Gravel Barrier-Lagoon Facies

Part of the rusty lithosome is composed of deposits which may represent a coastal environment. The coastal environment interdigitates with the braid-delta facies as well as the shoreface deposits of the shelf assemblage. In the strict sense of the definition of braid- and fan-deltas (McPherson *et al.*, 1988; Nemec and Steel, 1988), coastal settings compose part of the above-mentioned environments. For the purposes of this study, however, the coastal environment will be described separately in accordance with the facies assemblages of Ricketts and Evenchick (1991). However, it may be prudent to amalgamate these facies during regional facies mapping.

The transition from nonmarine to marine (i.e. braided channel system or alluvial plain) is commonly subtle and may not be clearly recognized in the stratigraphic profile. In these situations, it is necessary to rely mainly on facies associations to facilitate further interpretations. For example, within the rusty lithosome near the top of Tsatia Mountain (EPG-III, 1744 m), a relatively thin (< 1 m) recessive carbonaceous black shale to siltstone horizon is situated at the top of a coarsening-upward cycle of rusty weathered chert pebble- to cobble-conglomerate. The underlying conglomerate (EPG-III, 1710-1744 m) consists of relatively continuous (> 5 m) seaward-sloping (southerly), low-angle, parallel to non-parallel cross-stratified beds. These beds display crudely developed normal grading up foreset dip as well as a poorly defined down-dip (southerly) clast imbrication (Figure 8). Together, the conglomerate and carbonaceous shale units

Figure 8.

Simplified map of the study area showing the results of the paleocurrent study in the form of rose diagrams. Solid rose diagrams represent a-axis orientations in channel conglomerates of the submarine canyon/gully assemblage. Orientations of a-axes reflect a/b plane clast imbrication (i.e. a-axes parallel-to-flow). Horizontally striped rose diagrams represent a/b plane clast imbrication in barrier-bar conglomerates of the coastal assemblage. Vertically striped rose diagram represents foresets of ripple cross-stratification in the overbank facies of the prodelta-slope assemblage (EPG-V, at 76 m). Thick solid black lines represent the locations of relevant stratigraphic sections. Solid black lines lead to the locations where measurements for the appropriate rose diagrams were taken; otherwise, the centre of the rose represents the data collection site. Vm = vector mean, also represented by the line with the arrow. N = sample number.

may represent a marine gravel barrier bar juxtaposed to a lagoon formed on the landward side of the barrier system (c.f. Carter *et al.*, 1987; Ricketts and Evenchick, 1991). The gravel bars may have been partially emergent as indicated by the poorly-developed seaward-dipping imbrication which is typical of beach conglomerates (Ethridge and Wescott, 1984). The poorly developed imbrication in concert with the lack of disc-shaped clasts may indicate rapid progradation and hence, rapid burial of gravel bars; this would tend to inhibit the formation of well-developed beach deposits.

An alternative explanation of the carbonaceous shale involves abandonment of braided channels within the braid delta facies. Abandoned channels are accompanied by decreased energy levels leading to the deposition of siltstones and shales. Progressive colonization by plants may lead to the development of organic-rich beds (Rust and Koster, 1984). However, abandonment of channels typically produces fining-upward cycles. Since this was not observed under the carbonaceous shale, the lagoonal interpretation is favoured.

4.5 Significance of Rusty Weathering

Most of the locally occurring rusty weathered units in the northern Bowser Basin include coarse-grained fractions. In the study area, rusty units are predominantly located within the fan-delta and coastal assemblages as well as at the tops of the uppermost coarsening-upward cycles of the shelf assemblage on

Tsatia Mountain. An exception to this spatial weathering trend includes the laterally persistent Waterfall Conglomerate, which is described in the channel facies of the submarine canyon/gully assemblage, and marks the lower boundary of the shelf assemblage. Near the northwestern edge of Tsatia Mountain (EPG-VI, 202-216 m), this conglomerate displays a pervasive rusty colour which grades to grey southward, down the depositional axis.

Red beds occur in a variety of environments, including hot arid and semi-arid climates (Walker, 1976) as well as warm and humid climates, all of which may be restricted to within 30° of the equator (Walker, 1974; Turner, 1980). Some examples of environments containing red beds include alluvial fans, delta plains, deserts, and pebbly braided rivers. Turner (1980) has suggested that the type of depositional environment strongly influences the diagenetic history and subsequently the reddening process.

The main cause of the rusty colouring in red beds is the presence of hematite. Hematite is known to form from detrital iron hydroxides which are abundant in silt- and clay-sized fractions of fluvial environments with well-developed flood plain deposits (Turner, 1980). This process is particularly common in warm, moist environments (Van Houten, 1968). Another means of producing hematite includes the diagenetic intrastratal alteration and dissolution of detrital and authigenic iron-bearing clay minerals and silicates (e.g. hornblende, biotite, augite, olivine). These processes provide the necessary iron for the formation of authigenic hematite (Walker, 1976; Turner, 1980). Oxidizing

conditions accomplished through subaerial exposure and/or oxygenated groundwaters are a prerequisite for the diagenetic formation of hematite (Walker, 1976, Chandler, 1980; Turner, 1980). Support for this statement comes from the suggestion that red beds did not become significant until free oxygen began to accumulate in the atmosphere approximately 2000 million years ago (Turner, 1980). The process of hematite formation and preservation can continue for millions of years until all iron sources are used up; this can only be accomplished if oxidizing conditions remain (Walker, 1976).

It is often difficult to establish the exact mode of formation of hematite in red beds. In this study, petrographic analysis indicates that there is a paucity of iron-bearing minerals. In addition, the clay content in rusty weathered sandstones and conglomerates is extremely low. It is possible that the rusty weathering occurred at the expense of detrital iron-bearing clays and minerals during diagenesis. Regardless, it appears that favourable conditions for the formation of hematite (i.e. oxidizing environment) existed in specific depositional environments, all of which are interpreted to represent relatively shallow marine to possibly subaerially-exposed. Hence, the inferred depositional environment is consistent with the occurrence of localized rusty weathered units.

Chapter 5 Provenance

5.1 Introduction

Detailed descriptions of rocks within the study area (see Chapter 4) confirm earlier observations that chert is the predominant component within sandstones and conglomerates of the Bowser Lake Group in the northern part of the Bowser Basin; the most likely source suggested for the chert is the oceanic Cache Creek Terrane northeast of the basin (Souther and Armstrong, 1966; Eisebacher, 1974b, c, 1981, 1985; Eisebacher *et al.*, 1974; Tipper and Richards, 1976; Monger, 1977; Currie, 1984; Ricketts and Evenchick, 1991). This source is in accord with paleoflow patterns within the Bowser Lake Group, which indicate a northeasterly source. Furthermore, regional bedrock geology maps indicate that the Cache Creek Terrane is the likely source for significant amounts of chert-bearing strata.

Three different methods were used to test the Cache Creek source terrane hypothesis. The first two methods involved identification of the various lithologies within the study area, comparing them to strata of the Cache Creek Terrane. The first method involved microscopic analysis of clasts to detect any stratigraphic changes in composition and textures of the chert clasts within the Bowser succession. The second method involved the examination of vertical and lateral clast composition changes within measured sections by means of megascopic point counts of conglomerate outcrops. The final method was a paleocurrent study

based on measurements of a-axis imbrication in conglomerates as well as foresets in ripple cross-laminated sandstones.

A brief discussion of the gross stratigraphy of the Cache Creek Terrane as well as the results of the above-described tests are presented below.

5.2 Gross Stratigraphy of the Cache Creek Terrane

The Cache Creek terrane of north-central British Columbia is mainly composed of three successions: (1) the Early Mississippian to Permian Cache Creek Complex, (2) the Upper Triassic Kutcho Formation, and (3) the Upper Triassic Sinwa Formation and Lower Jurassic Inklin Formation (Gabrielse, 1991). The Cache Creek complex, capped by intercalated chert and greywacke, is composed mostly of radiolarian-bearing chert, mafic volcanic rocks, limestone, argillite, gabbro, diorite, and ultramafic rocks, is inferred to represent an oceanic environment (Gabrielse, 1991). The Kutcho Formation that overlies the Cache Creek Complex in the southeastern tip of the Cache Creek Terrane is a bimodal volcanic assemblage consisting of calc-alkaline basaltic to andesitic, and rhyodacitic to rhyolitic, volcanoclastics (Thorstad and Gabrielse, 1986). The Kutcho Formation represents an island arc environment formed on an oceanic basement (Gabrielse, 1991). The Sinwa Formation, which overlies the Kutcho Formation, is a discontinuous unit of marble and micritic limestone, 250 m thick. It is overlain by the Inklin Formation, a clastic succession of black shale to minor conglomerate which contains clasts from the underlying Kutcho and Sinwa

formations (Thorstad and Gabrielse, 1986). The Sinwa and Inklin formations may represent an intra-arc or forearc setting (Gabrielse, 1991).

Structural and stratigraphic relationships of the Cache Creek Terrane with Stikinia and the Bowser Basin are illustrated in the cross-section presented in Figure 9. The vergence of large-scale structures and movement indicated on the major faults clearly suggest a southwestward component of thrusting. The age of initial thrusting of this succession along the King Salmon Fault is reported to be Bajocian or older (Gabrielse, 1991).

5.3 Petrographic Examination of Clast Compositions

Basin-fill compositions are commonly used to help determine source terrane(s) (Dickinson *et al.*, 1983). Analyses of compositions from conglomerate clasts are preferred because sandstone compositions are susceptible to selective preservation of stable minerals through diagenesis, humid climatic conditions, and tectonic recycling (Dickie, 1989). For this study, emphasis was placed on the recognition of stratigraphic changes in the proportion of the various clast types based on petrographic examination of thin sections (see Appendix I for locations of thin section specimens).

The results suggest that there are only subtle changes in the proportion of clast types throughout the study area. For example, massive and radiolarian-bearing chert occur in similar proportions within the channel conglomerates of the submarine canyon/gully assemblage and the rusty lithosome. The huge volume of

Figure 9.

Schematic cross-section showing the structural style and stratigraphic relationships in the Cache Creek Terrane and Stikinia. Modified after Gabrielse (1991).

chert incorporated into the Bowser Lake Group reflects substantial unroofing of the source terrane. The only map unit containing significant chert-bearing strata north and northeast of the study area is the Cache Creek Complex of the Cache Creek Terrane. A provenance study of conglomerates within the Ashman Formation northeast of the present study area also suggested that the Cache Creek Complex was the source of chert for the conglomerates (Currie, 1984). The relative percentages of silty chert or altered volcanic grains as well as monocrystalline quartz and feldspar also remain constant throughout the succession. Detrital glauconite which occurs in low amounts (1-3%) in the shelf and prodelta-slope assemblages was not observed in any other environments within the study area. This may be a reflection of the depositional environments where marine conditions favoured the formation of glauconite. Perhaps the most significant modal grain percentage variation occurs in the shelf assemblage sandstones, where up to 30% of the sandstones are composed of chlorite-rich rock fragments. This is significant because no other environments contain such a high proportion of these fragments. This may reflect a period of unroofing of a volcanic source to the north, such as the Kutcho Formation within the Cache Creek Terrane, or may represent partial unroofing of an exposed area within volcanics of the Hazelton Group that bound the Bowser Lake Group to the north.

5.4 Megascopic point counts

Conglomerates within the Bowser Lake Group provide an excellent way to directly observe gross compositional changes throughout the stratigraphic record. A relatively quick way to quantify compositional changes was accomplished through megascopic point counts of conglomerates according to composition. For example, a point count study performed on Todagin Mountain corresponding to section EPG-I consisted of clasts divided into black, green, and grey chert; a fourth division included felsic volcanic clasts that contain feldspar phenocrysts. Wherever a conglomerate unit was encountered within the section, a 100-square grid was drawn directly on the outcrop face (parallel and perpendicular to bedding). The grid size was determined by the mean clast size, with identifications at 100 counts providing the basis for establishing relative abundances of the different clast types (described above), as well as the percentage of matrix. The relative abundance of each clast type was then normalized to 100%. Figure 10 presents the results of the point count analyses for section EPG-I, showing that there are no appreciable vertical trends within the measured section. A similar lateral study was conducted within the Waterfall Conglomerate on Tsatia Mountain. The clast types were divided in the same fashion as for the EPG-I study, and counts were performed approximately 200 m apart over the entire width of the conglomerate (approximately 2 to 3 km). The results of the Waterfall Conglomerate study were similar to the one above, showing no lateral trends along the conglomerate unit.

Figure 10.

Clast compositions from section EPG-I. Note that the relative abundance of clast types remains constant throughout the section.

5.5 Paleocurrents

A northern source area is indicated by the southward trend of paleocurrents, based on a study of imbrication in channel conglomerates of the submarine canyon/gully assemblage, as well as the direction of inclination of foresets in ripple cross stratification within the overbank facies of the prodelta-slope assemblage. As mentioned in the previous chapter, most clasts within the channel conglomerates show a/b plane imbrication, with a-axis orientations parallel to flow, and upcurrent dips. This type of fabric is similar to coarse-grained deposits associated with turbidites (Walker, 1984) as well as those of deep-sea conglomerates described by Hein (1984). Measurements were taken where weathering of conglomerates permitted nearly three-dimensional views of clasts. Only tabular clasts measuring greater than 1 cm in length (a-axis) were recorded. For each clast, the azimuth and plunge of the a-axis were measured, and a minimum of 50 clasts were measured at each station. After correcting for bed tilt due to folding associated with regional deformation, the results of each station were tabulated to construct an equal-area rose diagram. Figure 8 presents the results of this study, which indicate that the a-axes plunge predominantly to the north/northeast. This suggests that the paleoflow direction was to the south/southwest.

Rose diagrams from several areas display a bimodal pattern, with most clasts dipping upcurrent, and a minor component dipping downcurrent. This result may have been due to deposition of clasts onto low-amplitude wavy

bedforms (e.g. Hein, 1982). The relatively small number of clasts deviating from the dominant a-axis parallel-to-flow and displaying transverse-to-flow orientations (a/b plane still dipping upcurrent) may indicate that such clasts rolled along the bed after initial deposition (Hein, 1982).

Few paleoflow indicators were observed other than the clast imbrication described above. One locality within the overbank facies of the prodelta-slope assemblage (EPG-V, 076 m) contains well-preserved ripple cross-stratification with excellent three-dimensional exposure of the foresets. Although only 20 measurements were taken, the rose diagram in Figure 8 suggests that the paleocurrents were to the south/southwest, in agreement with the imbrication observations described above. These findings are consistent with previous studies which suggest that the main paleoflow direction was to the south (Eisbacher, 1974b, 1981; Ricketts and Evenchick, 1991).

Chapter 6 Depositional Synthesis

6.1 Discussion

In the northern part of the Bowser Basin, the Bowser Lake Group reflects a dynamic depositional system formed in response to the accretion of Stikinia and Cache Creek. The Bowser Basin is inferred to represent a foredeep formed by loading of oceanic rocks of the Cache Creek terrane that were thrust southwestward onto Stikinian crust. The oceanic detritus was shed southward into the Bowser foredeep, which may have initiated as the Early to Middle Jurassic Hazelton trough.

Within the study area, the Bowser Lake Group constitutes a conformable, coarsening-upward succession that represents a regression from a proximal submarine fan through a variety of coastal depositional systems. A paleogeographic reconstruction based on this study is illustrated in Figure 11. The model is consistent with, but more detailed than, the reconstruction proposed by Ricketts and Evenchick (1991; their Figure 7). Nearshore, coastal and shallow-marine deposits consist of braid deltas, gravel barrier-bars and lagoons, channel-mouth bars of the fan/braid-delta front, and fan-delta slope deposits. Gravels of fan/braid-deltas fed directly onto submarine slopes where they were funnelled into gullies and small canyons, some of which fed submarine fans. Slumps were common on the slope, and most likely initiated on the upper slopes where slope oversteepening from shelf sedimentation caused instabilities. Slump scars may

Figure 11.

Schematic illustration of the depositional model envisioned for the Bowser Lake Group, northern Bowser Basin. This is a more detailed version of the regional paleoenvironmental reconstruction presented by Ricketts and Evenchick (1991) for the northern part of the basin. This model includes the spatial location of slumps on the prodelta-slope, the relative positioning of shelf deposits, in addition to deposits of the fan-delta assemblage. The figure is not to scale. The main sediment transport direction is S/SW. The following is a list of the abbreviations used in the figure, including their meanings:

- UFsf = upper fan facies of the submarine fan assemblage
- Slps = slump facies of the prodelta-slope assemblage
- Sps = slope facies of the prodelta-slope assemblage
- Ops = overbank facies of the prodelta-slope assemblage
- Cscg = channel facies of the submarine canyon/gully assemblage
- Ns = nearshore facies of the shelf assemblage
- Is = inner shelf facies of the shelf assemblage
- Os = outer shelf facies of the shelf assemblage
- Bfd = braid-delta facies of the fan-delta assemblage
- FFfd = fan-delta front facies of the fan-delta assemblage
- FSfd = fan-delta slope facies of the fan-delta assemblage
- BFfd = braid-delta front facies of the fan-delta assemblage
- BSfd = braid-delta slope facies of the fan-delta assemblage
- GBLc = gravel-barrier bar-lagoon facies of the coastal assemblage

have provided conduits for the accumulation and funnelling of fan/braid-delta gravels (Ricketts, 1990). Where slopes were sufficiently stable, sand-sized overbank deposits from neighbouring gully channels accumulated. Inter-fan/braid-delta regions consist of cyclical coarsening-upward sequences representing outer or quiet inner shelf to shoreface environments. The shelf morphology was irregular as indicated by abrupt changes in thickness and lateral extent of facies. Thick shelf sections such as on Tsatia Mountain, may reflect proximity to a source of sedimentary detritus to the north.

This depositional model poses additional questions. First, how were the gravels transported across the shelf to the slope where they accumulated in canyons and gullies? Second, how wide was the sandy shelf? Third, what were the probable causes of the cyclicity of the shelf deposits? These questions will be addressed in the following discussions.

6.2 How were the gravels transported to the slope and how wide was the shelf?

It is difficult to explain how coarse sediments can be transported across a sand-rich shelf any appreciable distance (tens of kilometres scale). One alternative is to invoke a drop in sea level. During low stands, nonmarine to shallow marine deposits can be eroded and redeposited due to subaerial exposure and increased gradients (Cant and Hein, 1986; Haq, 1991). Channels and disconformities formed during sea level minima can be covered by subsequent transgressions (Cant and Hein, 1986). In the present example, this explanation is

rejected because no laterally continuous disconformities were observed in the study area beneath the conglomerates. One possible test to this conclusion is that coastal assemblage conglomerates and channel conglomerates of the submarine canyon/gully assemblage should be differently aged. Unfortunately, the conglomerates are barren of fossils and therefore, their relative ages could not be established.

An alternative hypothesis appeals to a narrow shelf width (<5 km) across which gravels from fan- and braid-deltas would pass directly into gullies on the slope, transported mainly by fast-flowing, noncohesive debris flows (Ricketts and Evenchick, 1991). Support for a narrow shelf width comes from study of the Waterfall Conglomerate on Tsatia Mountain and a similar study of a Gilbert-style fan-delta, 4 km west of Tsatia Mountain (Ricketts and Evenchick, 1991). The Waterfall Conglomerate, traced down its depositional axis for approximately 2 km, grades from a sheetlike rusty weathered and poorly sorted conglomerate (possibly representing fan/braid-delta front or slope deposits) to one which is typical of the channel facies of the submarine canyon/gully assemblage. A Gilbert-style fan delta with large conglomerate foresets occurs at the base of the rusty lithosome 4 km west of Tsatia Mountain. This conglomerate grades down its depositional axis in a fashion similar to the Waterfall Conglomerate over a 2-4 km distance (Figure 3 in Ricketts and Evenchick, 1991; B.D. Ricketts, pers. comm., 1992). The nature of the downslope transition of these two conglomerate units, in addition to the rusty weathering up depositional axis, suggest that the conglomerates originated in

relatively shallow marine to possibly subaerial conditions, and graded to the slope over a distance of no more than 3 or 4 km. This would indicate that the width of the predominantly sand-rich interfan shelves would not have exceeded 4 km.

6.3 What are the causes of shelf cyclicity?

Rapid and cyclical changes in sedimentation can result from (1) eustatic sea-level changes, (2) tectonically induced changes in the rate of sediment supply, and (3) basin subsidence (regional tectonics) (Cant and Hein, 1986; Haq, 1991). These factors are not necessarily independent and can act in concert to amplify or subdue relative sea-level changes (Haq, 1991).

It is difficult to decipher which factor was responsible for changing the relative sea-level and the corresponding signature within the shelf strata. If the coarsening-upward cycles resulted from increased sedimentation due to separate thrust events in the Cache Creek Terrane to the northeast, a change in lithologies among cycles might be expected (Cant and Hein, 1986), assuming that each thrust event tapped a different source area. However, this does not appear to be the case because the compositions of the shelf sandstones are relatively homogeneous. Possibly the area studied is too small to permit the recognition of lithological changes related to changing source areas.

Basin subsidence also influences sedimentation patterns within a developing basin. For example, continuous subsidence is recorded in passive continental margins due to thermal cooling (Cant and Hein, 1986). In this

scenario, the space available for the accumulation of sediments or accommodation as it is termed, is relatively constant (assuming a constant sea-level in this case). However, the Bowser Basin is inferred to have formed in a foredeep tectonic environment. This tectonically active environment would produce an erratic and cyclical subsidence pattern due to thrust loading of Stikinian crust by oceanic rocks of the Cache Creek Terrane (Beaumont, 1981). In light of the previous paragraph, it is difficult to ascribe the coarsening-upward cycles to subsidence alone.

Finally, attempts have been made to compare the shelf strata cycles to published eustatic sea-level curves (Haq *et al.*, 1987). Using the terminology of sequence-stratigraphy, the coarsening-upward cycles of the Shelf Assemblage represent parasequences. A parasequence is composed of a conformable succession of genetically related beds or bedsets which are bounded above and below by marine flooding surfaces (Haq, 1991). Unfortunately, the resolution of the eustatic sea-level curves presented by Haq *et al.* (1987) is not high enough to discern parasequences within the shelf assemblage. However, the age of the shelf assemblage, based on biostratigraphy, ranges from Early to Late Callovian and possibly Early Oxfordian. According to published curves, sea-levels rose during the Callovian and continued throughout the Oxfordian (Haq *et al.*, 1987). The age range for the entire succession within the study area is Late Bathonian to Middle and perhaps Late Oxfordian. The sea-level curves indicate that sea-levels were relatively low in the Late Bathonian and steadily rose until they reached a

peak in the Kimmeridgian. This is interesting because the Bowser Lake Group represents an overall coarsening-upward and shoaling of the lithofacies. In an environment with steadily rising sea levels one would expect to find an overall transgressive succession. One of the main contributing factors to the coarsening-upward trend in this study might include high sedimentation rates. This is supported by the fact that an abundance of coarse-grained detritus (primarily chert pebbles and cobbles) was shed into the basin during this time. In any event, the depositional environment of the Bowser Lake Group reflects high energies and relatively steep gradients.

It appears that the sedimentary pattern of the Bowser Lake Group is quite complex, and cannot be explained by any one of the factors discussed above. The more likely situation is that all three forces interacted to produce the preserved stratigraphic succession.

6.4 Conclusions

The Bowser Lake Group in the northern Bowser Basin is inferred to represent an overlap assemblage, recording a tectonic link between the volcanic rocks of Stikinia and the oceanic Cache Creek Terrane. The main purpose of this investigation was to test the regional sedimentological model proposed by Ricketts (1990) and Ricketts and Evenchick (1991) for the northern part of the basin. This was accomplished through a detailed sedimentological study in a structurally uncomplicated area at the northern limit of the basin. The results of this study

are consistent with Ricketts' model; however, additional details for some assemblages have been added. Most of the facies assemblages and lithofacies described by Ricketts and Evenchick (1991) are documented within the immediate study area (see Table 1). The deep basin assemblage is the only assemblage missing, but does underlie other parts of the basin. The fan-delta assemblage has been modified to include a fan/braid-delta front, and fan/braid-delta slope facies. The Gilbert delta facies was not observed.

The presence of rusty weathering in conglomerates and sandstones is suggestive of oxidizing conditions during deposition and diagenesis, and may indicate shallow-marine to subaerial exposure. The paleoenvironmental reconstruction of the study area depicts a shelf width of no more than 4 km. This inference is based on observations of the transition of coastal and shallow marine conglomerates to those typical of the submarine canyon/gully assemblage over a distance of 2 to 4 km. The cyclicity within the shelf assemblage, in addition to drowning of the rusty lithosome by the shelf assemblage, may reflect sea-level changes induced by a combination of increased sedimentation and changes in basin subsidence related to tectonism during formation of the Bowser Basin foredeep. Eustatic sea-level changes may also account for some of the cyclicity. The overall coarsening-upward and shoaling of the Bowser Lake Group within the study area is the opposite of what should be expected in comparison with the continuous long-term rise in sea-levels reported for the Late Bathonian through Oxfordian (Haq *et al.*, 1987). High rates of sedimentation, which accords with the

coarse-grained nature of the Bowser Lake Group strata, may account for the coarsening-upward pattern.

A three-part provenance investigation included a study of imbrication in the conglomerates, megascopic point counts of the conglomerates, and petrographic analyses of thin sections from rock specimens. The imbrication study, supported by other paleoflow indicators, suggests that the main paleoflow direction was to the south/southwest. Megascopic point counts of conglomerates failed to reveal any lateral or vertical compositional trends, and petrographic examination also indicated that the relative proportions of the various components remain constant laterally and vertically. An anomalous increase in chlorite-rich rock fragments in sandstones of the shelf assemblage may suggest partial unroofing of a volcanic source to the north. Based on regional geology maps and the results of the provenance study, the Cache Creek Terrane north and northeast of the study area is the most likely source for sediments of the Bowser Lake Group. This is in agreement with previous studies.

Although an understanding of the stratigraphic framework of the Bowser Lake Group has been significantly improved through the Bowser Basin project, numerous problems still need to be addressed. For example, it is not known whether the coarsening-upward succession, observed within the study area, is repeated higher in the stratigraphy or whether the reappearance of fine-grained rocks further to the south is due solely to structural deformation. This has important ramifications for interpretations of basin evolution as well as delineating regionally mappable units. Secondly, a provenance study at a larger scale, both vertically and laterally, may reveal broad clast composition patterns related to unroofing of the Cache Creek Terrane to the northeast. Finally, the channel conglomerates of the submarine canyon/gully assemblage provide an excellent laboratory to study the development of imbrication fabrics in conglomerates.

REFERENCES

- Aigner, T. 1985. Storm depositional systems. Dynamic stratigraphy in modern and ancient shallow-marine sequences. *Lecture Notes in Earth Sciences*, v. 3, 174 pp.
- Anderson, R.G., and Thorkelson, D.J. 1990. Mesozoic stratigraphy and setting for some mineral deposits in Iskut River map area, northwestern British Columbia. *In* Current Research, Part E. Geological Survey of Canada, Paper 90-1E, pp. 131-139.
- Atkinson, C.D., Goesten, M.J.B.G., Speksnijder, A., and van der Vlugt, W. 1986. Storm-generated sandstone in the Miocene Miri Formation, Seria Field, Brunei (N.W. Borneo). *In* Shelf sands and sandstones. *Edited by* R.J. Knight and J.R. McLean. Canadian Society of Petroleum Geologists, Memoir 11, pp. 213-240.
- Beaumont, C. 1981. Foreland basins. *Geophysical Journal of the Royal Astronomical Society*, 65: 291-329.
- Bouma, A.H., Berryhill, H.L., Knebel, H.J., and Brenner, R.L. 1982. Continental shelf and epicontinental seaways. *In* Sandstone depositional environments. *Edited by* P.A. Scholle and D. Spearing. American Association of Petroleum Geologists, Memoir 31, pp. 281-327.
- Brenchley, P.J., Romano, M., and Gutierrez Marco, J.C. 1986. Proximal and distal hummocky cross-stratified facies on a wide Ordovician shelf in Iberia. *In* Shelf sands and sandstones. *Edited by* R.J. Knight and J.R. McLean. Canadian Society of Petroleum Geologists, Memoir 11, pp. 241-255.
- Buckham, A.F., and Latour, B.A. 1950. The Groundhog coalfield, British Columbia. Geological Survey of Canada, Bulletin 16.
- Bull, W.B. 1972. Recognition of alluvial-fan deposits in the stratigraphic record. *In* Recognition of ancient sedimentary environments. *Edited by* J.K. Rigby and W.K. Hamblin. Society of Economic Paleontologists and Mineralogists, Special Publication 16, pp. 63-83.
- Bull, W.B. 1977. The alluvial-fan environment. *Progress in Physical Geography*, 1: 222-270.

- Bustin, R.M. 1984. Coalification levels and their significance in the Groundhog coalfield, north-central British Columbia. *International Journal of Coal Geology*, 4: 21-44.
- Bustin, R.M., and Moffat, I. 1983. Groundhog coalfield, central British Columbia: Reconnaissance stratigraphy and structure. *Bulletin of Canadian Petroleum Geology*, 31: 231-245.
- Cant, D.J., and Hein, F.J. 1986. Depositional sequences in ancient shelf sediments: some contrasts in style. *In Shelf sands and sandstones. Edited by R.J. Knight and J.R. McLean. Canadian Society of Petroleum Geologists, Memoir 11, pp. 303-312.*
- Carter, R.W.G., Orford, J.D., Forbes, D.L., and Taylor, R.B. 1987. Gravel barriers, headlands and lagoons: an evolutionary model. *In Coastal sediments '87. Edited by N.C. Kraus, Proceedings, pp. 1776-1792.*
- Chandler, F.W. 1980. Proterozoic redbed sequences of Canada. *Geological Survey of Canada, Bulletin 311.*
- Colmenaro, J.R., Agueda, J.A., Fernandez, L.P., Salvador, C.I., Bahamonde, J.R., and Barba, P. 1988. A coal-bearing fan-delta complex in the Westphalian D of the Central Coal Basin, Cantabrian Mountains, northwestern Spain: implications for the recognition of humid-type fan deltas. *In Fan deltas. Edited by W. Nemec and R.J. Steel. Blackie and Son Ltd., London, pp. 286-302.*
- Coney, P., Jones, D.L., and Monger, J.W.H. 1980. Cordilleran suspect terranes. *Nature*, 288: 329-333.
- Cookenboo, H.O., and Bustin, R.M. 1989. Jura-Cretaceous (Oxfordian to Cenomanian) stratigraphy of the north-central Bowser Basin, northern British Columbia. *Canadian Journal of Earth Sciences*, 26: 1001-1012.
- _____. 1990. Lithostratigraphy of the northern Skeena Mountains, British Columbia. *In Current Research, Part F. Geological Survey of Canada, Paper 90-1F, pp. 151-156.*
- Cordey, F., Mortimer, N., DeWever, P., and Monger, J.W.H. 1987. Significance of Jurassic radiolarians from the Cache Creek terrane, British Columbia. *Geology*, 15: 1151-1154.

- Cordey, F., Gordey, S.P., and Orchard, M.J. 1991. New biostratigraphic data for the northern Cache Creek Terrane, Teslin map area, southern Yukon. *In* Current Research, Part E. Geological Survey of Canada, Paper 91-1E, pp. 67-76.
- Currie, L.D. 1984. The provenance of chert clasts in the Ashman conglomerates of the northeastern Bowser Basin. BSc. thesis, Queens University, Kingston, Ontario.
- Davis, R.A. Jr. 1983. Depositional systems: a genetic approach to sedimentary geology. Prentice-Hall, Inc., New Jersey.
- DeCelies, P.G. 1987. Variable preservation of middle Tertiary, coarse-grained, nearshore to outer-shelf storm deposits in southern California. *Journal of Sedimentary Petrology*, 57: 250-264.
- Dickie, J.R. 1989. Sedimentary Response to Arc-Continent Transpressive Tectonics, Laberge Conglomerates (Jurassic), Whitehorse Trough, Yukon Territory. MSc. thesis, Dalhousie University, Halifax, Nova Scotia, 361 pp.
- Dickinson, W.R., Beard, L.S., Brakenridge, G.R., Erjavec, J.L., Ferguson, R.C., Inman, K.F., Knepp, R.A., Lindberg, F.A., and Ryberg, P.T. 1983. Provenance of North American Phanerozoic sandstones in relation to tectonic setting. *Geological Society of America Bulletin*, 94: 222-235.
- Dott, R.H., Jr., and Bourgeois, J. 1982. Hummocky stratification: sequences. *Geological Society of America, Bulletin*, 93: 663-680.
- Duffell, S., and Souther, J.G. 1964. Geology of Terrace map-area, British Columbia. Geological Survey of Canada, Memoir 329.
- Eisbacher, G.H. 1974a. Sedimentary history and tectonic evolution of the Sustut and Sifton basins, north-central British Columbia. Geological Survey of Canada, Paper 73-31.
- _____. 1974b. Deltaic sedimentation in the northeastern Bowser Basin, British Columbia. Geological Survey of Canada, Paper 73-33.
- _____. 1974c. Evolution of successor basins in the Canadian Cordillera. *In* Modern and ancient geosynclinal sedimentation. *Edited by* R.H. Dott, Jr. and R.H. Shaver. Society of Economic Paleontologists and Mineralogists, Special Publication 19, pp. 274-291.

- _____. 1981. Late Mesozoic-Paleogene Bowser Basin molasse and Cordilleran tectonics, western Canada. *In* Sedimentation and tectonics in alluvial basins. *Edited by* A.D. Miall. Geological Association of Canada, Special Paper 23, pp. 125-151.
- _____. 1985. Pericollisional strike-slip faults of synorogenic basins, Canadian Cordillera. *In* Strike-slip deformation, basin formation, and sedimentation. *Edited by* K.T. Biddle and N. Christie-Blick. Society of Economic Paleontologists and Mineralogists, Special Publication 37, pp. 265-282.
- Eisbacher, G.H., Carrigy, M.A., and Campbell, R.B. 1974. Paleodrainage patterns and late-orogenic basins of the Canadian Cordillera. *In* Tectonics and sedimentation. *Edited by* W.R. Dickinson. Society of Economic Paleontologists and Mineralogists, Special Publication 22, pp. 143-166.
- Ethridge, F.G., and Wescott, W.A. 1984. Tectonic setting, recognition and hydrocarbon reservoir potential of fan-delta deposits. *In* Sedimentology of gravels and conglomerates. *Edited by* E.H. Koster and R.J. Steel. Canadian Society of Petroleum Geologists, Memoir 10, pp. 217-235.
- Evenchick, C.A. 1986. Structural style of the northeast margin of the Bowser Basin, Spatsizi map area, north-central British Columbia. *In* Current Research, Part B. Geological Survey of Canada, Paper 86-1B, pp. 733-730.
- _____. 1987. Stratigraphy and structure of the northeast margin of the Bowser Basin, Spatsizi map area, north-central British Columbia. *In* Current Research, Part A. Geological Survey of Canada, Paper 87-1A, pp. 719-726.
- _____. 1988. Structural style and stratigraphy in northeast Bowser and Sustut basins, north-central British Columbia. *In* Current Research, Part E. Geological Survey of Canada, Paper 88-1E, pp. 91-95.
- _____. 1989. Stratigraphy and structure in east Spatsizi map area, north-central British Columbia. *In* Current Research, Part E. Geological Survey of Canada, Paper 89-1E, pp. 133-138.
- _____. 1991a. Jurassic stratigraphy of east Telegraph Creek and West Spatsizi map area, British Columbia. *In* Current Research, Part A. Geological Survey of Canada, Paper 91-1A, pp. 155-162.
- _____. 1991b. Geometry, evolution, and tectonic framework of the Skeena fold belt, north-central British Columbia. *Tectonics*, 10: 527-546.

- _____. 1991c. Structural relationships of the Skeena Fold Belt west of the Bowser Basin, northwest British Columbia. *Canadian Journal of Earth Sciences*, 28: 973-983.
- Evenchick, C.A., and Green, G.M. 1990. Structural style and stratigraphy of southwest Spatsizi map area, British Columbia. *In* Current Research, Part F. Geological Survey of Canada, Paper 90-1F, pp. 135-144.
- Gabrielse, H. 1991. Late Paleozoic and Mesozoic terrane interactions in north-central British Columbia. *Canadian Journal of Earth Sciences*, 28: 947-957.
- Gabrielse, H., and Tipper, H.W. 1984. Bedrock geology of Spatsizi map area (104H). Geological Survey of Canada, Open File 1005.
- Gabrielse, H., and Yorath, C.J. 1989. DNAG#4. The Cordilleran Orogen in Canada. *Geoscience Canada*, 16: 67-83.
- Geological Survey of Canada. 1957. Stikine River Area, Cassiar District, British Columbia. Geological Survey of Canada, Map 9-1957.
- Gloppen, T.G., and Steel, R.J. 1981. The deposits, internal structure and geometry in six alluvial fan-fan delta bodies (Devonian-Norway) - A study in the significance of bedding sequence in conglomerate. *In* Recent and ancient nonmarine depositional environments: models for exploration. *Edited by* F.G. Ethridge and R.M. Flores. Society of Economic Paleontologists and Mineralogists, Special Publication 31, pp. 49-69.
- Green, G.M. 1991. Detailed sedimentology of the Bowser Lake Group, northern Bowser Basin, British Columbia. *In* Current Research, Part A. Geological Survey of Canada, Paper 91-1A, pp. 187-195.
- Greenwood, B., and Sherman, D.J. 1986. Hummocky cross-stratification in the surf zone: flow parameters and bedding genesis. *Sedimentology*, 33: 33-45.
- Greig, C.J. 1991. Jurassic stratigraphy of east Telegraph Creek and West Spatsizi map area, British Columbia. *In* Current Research, Part A. Geological Survey of Canada, Paper 91-1A, pp. 197-205.
- Greig, C.J. 1992. Fieldwork in the Oweegee and Snowslide ranges and Kinskuch Lake area, northwestern British Columbia. *In* Current Research, Part A. Geological Survey of Canada, Paper 92-1A, pp. 145-155.

- Haq, B.U. 1991. Sequence stratigraphy, sea-level change, and significance for the deep sea. *In* Sedimentation, tectonics and eustasy: sea-level changes at active margins. Special Publications, International Association of Sedimentologists, 12: 3-39.
- Haq, B.U., Hardenbol, J., and Vail, P.R. 1987. Chronology of fluctuating sea levels since the Triassic. *Science*, 235: 1156-1167.
- Harms, J.C., Southard, J., Spearing, D.R., and Walker, R.G. 1975. Depositional environments as interpreted from primary sedimentary structures and stratification sequences. Society of Economic Paleontologists and Mineralogists, Dallas, Texas, Short Course No. 2, 161 pp.
- Hein, F.J. 1982. Depositional mechanisms of deep-sea coarse clastic sediments, Cap Enrage Formation, Quebec. *Canadian Journal of Earth Sciences*, 19: 267-287.
- 1984. Deep sea and fluvial braided channel conglomerates: a comparison of two case studies. *In* Sedimentology of gravels and conglomerates. *Edited by* E.H. Koster and R.J. Steel. Canadian Society of Petroleum Geologists, Memoir 10, pp. 33-49.
- Helwig, T. 1970. Slump folds and early structures, northeastern Newfoundland Appalachians. *Journal of Geology*, 78: 172-187.
- Hey, M.H. 1954. A new review of the chlorites. *Mineralogy Magazine*, 30: 277 pp.
- Hooke, R.L. 1967. Processes on arid-region alluvial fans. *Journal of Geology*, 75: 438-460.
- Howell, D.G., and Normark, W.R. 1982. Sedimentology of submarine fans. *In* Sandstone depositional environments. *Edited by* P.A. Scholle and D. Spearing. American Association of Petroleum Geologists, Memoir 31, pp. 365-404.
- Ineson, J.R. 1985. Submarine glide blocks from the Lower Cretaceous of the Antarctic Peninsula. *Sedimentology*, 32: 659-670.
- Irving, E., and Thorkelson, D.J. 1990. On determining paleohorizontal and latitudinal shifts: paleomagnetism of Spences Bridge Group, British Columbia. *Journal of Geophysical Research*, 95: 213-234.

- Irving, E., Monger, J.W.H., and Yole, R.W. 1980. New paleomagnetic evidence for displaced terranes in British Columbia. *In* The continental crust and its mineral deposits. Geological Association of Canada, Special Paper 20, pp. 444-456.
- Ito, M, and Masudo, F. 1988. Late Cenozoic deep-sea to fan-delta sedimentation in an arc-arc collision zone, central Honshu, Japan: sedimentary response to varying plate-tectonic regime. *In* Fan deltas. *Edited by* W. Nemec and R.J. Steel. Blackie and Son Ltd., London, pp. 400-418.
- Johnson, H.D., and Baldwin, C.T. 1986. Shallow siliciclastic seas. *In* Sedimentary environments and facies, 2nd Edition. *Edited by* H.G. Reading. Blackwell Scientific Publications, Boston, pp. 229-282.
- Kaip, A.W. 1991. Synsedimentary deformation and depositional environment of the prodelta-slope assemblage, Ashman Formation, Bowser Lake Group, north-central British Columbia. B.Sc. thesis, Carleton University, Ottawa, Ontario, 44 pp.
- Kleinspehn, K.L., Steel, R.J., Johannessen, E., and Netland, A. 1984. Conglomeratic fan-delta sequences, Late Carboniferous-Early Permian, western Spitsbergen. *In* Sedimentology of gravels and conglomerates. *Edited by* E.H. Koster and R.J. Steel. Canadian Society of Petroleum Geologists, Memoir 10, pp. 279-294.
- Koo, J. 1986. Geology of the Klappan coalfield in northwestern British Columbia (104H/2,3,6,7). *In* Geological fieldwork, 1985. British Columbia Ministry of Energy Mines and Petroleum Resources, Paper 1986-1, pp. 225-228.
- Koster, E.H., and Steel, R.J. (ed.). 1984. Sedimentology of gravels and conglomerates. Canadian Society of Petroleum Geologists, Memoir 10, 441 pp.
- Lewis, D. 1971. Slumping on a continental slope inclined at 1-4°. *Journal of Sedimentology*, 16: 97-110.
- . 1980. Storm-generated graded beds and debris flow deposits with *Ophiomorpha* in a shallow offshore sequence at Nelson, South Island, New Zealand. *Journal of Geology and Geophysics*, 23: 353-369.
- Lowe, D.R. 1979. Sediment gravity flows: their classification and some problems of application to natural flows and deposits. *In* Geology of continental slopes. *Edited by* L.J. Doyle and D.H. Pilkey. Society of Economic Paleontologists and Mineralogists, Special Publication 27, pp. 75-82.

- _____. 1982. Sediment gravity flows: II depositional models with special reference to the deposits of high density turbidity currents. *Journal of Sedimentary Petrology*, 52: 279-297.
- MacLeod, S.E., and Hills, L.V. 1990. Conformable Late Jurassic (Oxfordian) to Early Cretaceous strata, northern Bowser Basin, British Columbia: a sedimentological and paleontological model. *Canadian Journal of Earth Sciences*, 27: 988-998.
- Malloch, G.S. 1914. The Groundhog coal field, British Columbia. Geological Survey of Canada Summary Report, 1912, pp. 69-101.
- Marsden, H., and Thorkelson, D.J. 1992. Geology of the Hazelton volcanic belt in British Columbia: implications for the Early to Middle Jurassic evolution of Stikinia. *Tectonics*, in press.
- Marzo, M., and Anadon, P. 1988. Anatomy of a conglomeratic fan-delta complex: the Eocene Montserrat Conglomerate, Ebro Basin, northeastern Spain. *In Fan deltas. Edited by W. Nemec and R.J. Steel. Blackie and Son Ltd., London*, pp. 318-340.
- McPherson, R.J., Shanmugam, G., and Moiola, R. 1988. Fan-deltas and braid-deltas: conceptual problems. *In Fan deltas. Edited by W. Nemec and R.J. Steel. Blackie and Son Ltd., London*, pp. 14-22.
- Miall, A.D. 1977. A review of the braided-river depositional environment. *Earth Science Reviews*, 13: 1-62.
- _____. 1978. Lithofacies types and vertical profile models in braided river deposits: a summary. *In Fluvial sedimentology. Edited by A.D. Miall. Canadian Society of Petroleum Geologists, Memoir 5*, pp. 597-604.
- Middleton, G.V., and Hampton, M.A. 1973. Sediment gravity flows: mechanics of flow and deposition. *In Turbidites and deep-water sediments. Society of Economic Paleontologists and Mineralogists, Pacific Section, Los Angeles, Short Course*, 38 pp.
- _____. 1976. Subaqueous sediment transport and deposition by sediment gravity flows. *In Marine transport and environmental management. Edited by D.G. Stanley and D.J. Swift. John Wiley, New York*, pp. 197-217.

- Moffat, I.W., Bustin, R.M., and Rouse, G.E. 1988. Biochronology of selected Bowser Basin strata; tectonic significance. *Canadian Journal of Earth Sciences*, 25: 1571-1578.
- Monger, J.W.H. 1977. Upper Paleozoic rocks of the western Canadian Cordillera and their bearing on Cordilleran evolution. *Canadian Journal of Earth Sciences*, 14: 1832-1859.
- _____. 1984. Cordilleran tectonics: a Canadian perspective. *Bulletin of the Geological Society of France*, 7: 255-278.
- _____. 1989. Overview of Cordilleran geology. *In Western Canada Sedimentary Basin: A Case History. Edited by B.D. Ricketts. Canadian Society of Petroleum Geologists*, P. 9-32.
- Monger, J.W.H., and Irving, E. 1980. Northward displacement of north-central British Columbia. *Nature*, 285: 289-294.
- Monger, J.W.H., and Ross, C.A. 1971. Distribution of fusuliniaceans in the Canadian Cordillera. *Canadian Journal of Earth Sciences*, 8: 259-278.
- Monger, J.W.H., Price, R.A., and Tempelman-Kluit, D. 1982. Tectonic accretion and the origin of the two major metamorphic and plutonic belts in the Canadian Cordillera. *Geology*, 10: 70-75.
- Mustard, P.S. 1990. Upper Proterozoic-Lower Cambrian sedimentary rocks of the Mount Harper Group, Ogilvie Mountains, Yukon. Ph.d. thesis, Carleton University, Ottawa, Ontario, 333 pp.
- Mount, J.F. 1984. Mixing of siliciclastic and carbonate sediments in shallow shelf environments. *Geology*, 12: 432-435.
- _____. 1985. Mixed siliciclastic and carbonate sediments: a proposed first-order textural and compositional classification. *Sedimentology*, 32: 435-442.
- Mutti, E., and Ricci Lucci, F. 1972. Turbidity of the northern Apennines: introduction to facies analysis. (English translation by T.H. Nilsen, 1978). *International Geological Reviews*, 20: 125-166.

- Nardin, T.R., Hein, F.J., Garsline, D.S., and Edwards, B.D. 1979. A review of mass movement processes, sediment and acoustic characteristics, and contrasts in slope and base-of-slope systems versus canyon-fan-basin floor systems. *In* The geology of continental slopes. *Edited by* L.J. Doyle and O.H. Pilkey. Society of Economic Paleontologists and Mineralogists, Special Publication 27, pp. 61-73.
- Nelson, C.H. 1982. Modern shallow-water graded sand layers from storm surges, Bering shelf: a mimic of Bouma sequences and turbidite systems. *Journal of Sedimentary Petrology*, 52: 537-545.
- Nemec, W., and Steel, R.J. 1984. Alluvial and coastal conglomerates: their significant features and some comments on gravelly mass-flow deposits. *In* Sedimentology of gravels and conglomerates. *Edited by* E.H. Koster and R.J. Steel. Canadian Society of Petroleum Geologists, Memoir 10, pp.1-31.
- _____. 1988. What is a fan-delta and how do we recognize it? *In* Fan deltas. *Edited by* W. Nemec and R.J. Steel. Blackie and Son Ltd., London, pp. 1-13.
- Nilsen, T.H. 1982. Alluvial fan deposits. *In* Sandstone depositional environments. *Edited by* P.A. Scholle and D.R. Spearing. American Association of Petroleum Geologists, Memoir 21, pp. 49-86.
- Postma, G. 1984a. Slumps and their deposits in the fan delta front and slope. *Geology*, 12: 27-30.
- _____. 1984b. Mass-flow conglomerates in a submarine canyon: Abrioja fan-delta, Pliocene, southeast Spain. *In* Sedimentology of gravels and conglomerates. *Edited by* E.H. Koster and R.J. Steel. Canadian Society of Petroleum Geologists, Memoir 10, pp. 237-258.
- Poulton, T.P., Callomon, J.H., and Hall, R.J. 1991. Bathonian through Oxfordian (Middle and Upper Jurassic) marine macrofossil assemblage and correlations, Bowser Lake Group, west-central Spatsizi map area, northwestern British Columbia. *In* Current Research, Part A. Geological Survey of Canada, Paper 91-1A, pp. 59-63.
- Read, P.B. and Psutka, J.F. 1990. Geology, Falue Lake east-half (104H/13E) and Cullivan Creek (104H/14) map areas. Geological Survey of Canada, Open File 2241.

- Reinson, G.E. 1984. Barrier island and associated strand-plain systems. *In* Facies models. 2nd edition. *Edited by* R.G. Walker. Geoscience Canada Reprint Series 1, Geological Survey of Canada Publication, pp. 119-140.
- Ricci Lucchi, F. 1975. Depositional cycles in two turbidite formations of northern Apennines (Italy). *Journal of Sedimentary Petrology*, 45: 3-43.
- Richards, M.T. 1986. Tidal bed form migration in shallow marine environments: evidence from the Lower Triassic, western Alps, France. *In* Shelf sands and sandstones. *Edited by* R.J. Knight and J.R. McLean. Canadian Society of Petroleum Geologists, Memoir 11, pp. 257-276.
- Richards, T.A., and Gilchrist, R.D. 1979. The Groundhog coal area, British Columbia. *In* Current Research, Part B. Geological Survey of Canada, Paper 79-1B, pp. 411-414.
- Ricketts, B.D. 1990. A preliminary account of sedimentation in the lower Bowser Lake Group, northern British Columbia. *In* Current Research, Part F. Geological Survey of Canada, Paper 90-1F, pp. 145-150.
- Ricketts, B.D., and Evenchick, C.A. 1991. Analysis of the Middle to Upper Jurassic Bowser Basin, northern British Columbia. *In* Current Research, Part A. Geological Survey of Canada, Paper 91-1A, pp. 65-74.
- Ricketts, B.D., Evenchick, C.A., Murphy, D.C., and Anderson, R.G. in press. Bowser Basin, northern British Columbia: constraints on the timing of initial subsidence and Stikinia/North America terrane interactions. *Geology*.
- Rupke, N.A. 1986. Deep clastic seas. *In* Sedimentary environments and facies. 2nd edition. *Edited by* H.G. Reading. Blackwell Scientific Publications, Boston, pp. 372-415.
- Rust, B.R. 1978. Depositional models for braided alluvium. *In* Fluvial sedimentology. *Edited by* A.D. Miall. Canadian Society of Petroleum Geologists, Memoir 5, pp. 605-624.
- Rust, B.R., and Koster, E.H. 1984. Coarse alluvial deposits. *In* Facies models. 2nd edition. *Edited by* R.G. Walker. Geoscience Canada Reprint Series 1, Geological Association of Canada Publication, pp. 53-70.
- Shanmugam, G., and Moiola, R.J. 1988. Submarine fans: characteristics, models, classification and reservoir potential. *Earth Science Reviews*, 24: 383-428.

- Smith, P.L., and Tipper, H.W. 1986. Plate tectonics and paleobiogeography: Early Jurassic (Pliensbachian) endemism and diversity. *Palaios*, 1: 399-412.
- Smith, P.L., Thomson, R.C., and Tipper, H.W. 1984. Lower and Middle Jurassic sediments and volcanics of the Spatsizi map area, British Columbia. *In* Current Research, Part A. Geological Survey of Canada, Paper 84-1A, pp.117-120.
- Souther, J.G. 1971. Geology and mineral deposits of Tulsequah map-area, British Columbia. Geological Survey of Canada, Memoir 362, pp. 1-84.
- _____. 1972. Telegraph Creek map-area. Geological Survey of Canada, Paper 71-44, pp. 1-38.
- _____. in press. Maitland, Canada. *In* Volcanoes of North America. *Compiled by* C.A. Wood and J. Kienle. Cambridge University Press, New York.
- Souther, J.G., and Armstrong, J.E. 1966. North Central Belt of the Cordillera of British Columbia. Canadian Institute of Mining and Metallurgy, Special Volume 8, pp. 171-184.
- Swift, D.J.P., Figueiredo, A.G., Jr., Freeland, G.L., and Ortel, G.F. 1983. Hummocky cross-stratification and megaripples: a geological double standard? *Journal of Sedimentary Petrology*, 53: 1295-1317.
- Swift, D.J.P., Han, G., and Vincent, C.E. 1986. Fluid processes and sea-floor response on a modern storm-dominated shelf: Middle Atlantic shelf of North America. Part 1: The storm-current regime. *In* Shelf sands and sandstones. *Edited by* R.J. Knight and J.R. McLean. Canadian Society of Petroleum Geologists, Memoir 11, pp. 99-119.
- Thomson, R.C., Smith, P.L., and Tipper, H.W. 1986. Lower to Middle Jurassic (Pliensbachian to Bajocian) stratigraphy of the northern Spatsizi area, north-central British Columbia. *Canadian Journal of Earth Sciences*, 23: 1963-1973.
- Thorkelson, D.J. 1988. Jurassic and Triassic volcanic and sedimentary rocks in the Spatsizi map-area, north-central British Columbia. *In* Current Research, Part E. Geological Survey of Canada, Paper 88-1E, pp. 43-48.
- Thorstad, L.E., and Gabrielse, H. 1986. The Upper Triassic Kutcho Formation, Cassiar Mountains, north-central British Columbia. Geological Survey of Canada, Paper 86-16, pp. 1-53.

- Tipper, H.W., and Richards, T.A. 1976. Jurassic stratigraphy and history of north-central British Columbia. Geological Survey of Canada, Bulletin 270, 73 p.
- Turner, P. 1980. Continental red beds. *In* Developments in sedimentology, 29. Elsevier Scientific Publishing Company, New York, 562 pp.
- Van Houten, F.B. 1968. Iron oxides in red beds. Geological Society of America Bulletin, 79: 399-416.
- Walker, T.R. 1974. Formation of red beds in moist tropical climates: a hypothesis. Geological Society of America Bulletin, 85: 633-638.
- _____. 1976. Diagenetic origin of continental red beds. *In* The continental Permian in central, west, and south Europe. Edited by H. Falke. D. Reidel, Dordrecht-Holland, pp. 240-282.
- Walker, R.G. 1984. Shelf and shallow marine sands. *In* Facies models. 2nd edition. Edited by R.G. Walker. Geoscience Canada Reprint Series 1, Geological Survey of Canada Publication, pp. 141-170.
- Wheeler, J.O., and McFeely, P. (compilers) 1987. Tectonic assemblage map of the Canadian Cordillera and adjacent parts of the United States of America. Geological Survey of Canada, Open File 1565.
- Wheeler, J.O., Brookfield, A.J., Gabrielse, H., Monger, J.W.H., Tipper, H.W., and Woodsworth, G.J. (comp.) 1991. Terrane map of the Canadian Cordillera. Geological Survey of Canada, Map 1713A.
- Woodcock, N.H. 1979. Sizes of submarine slides and their significance. Journal of Structural Geology, 1: 137-142.
- Wright, L.D. 1977. Sediment transport and deposition at river mouth: a synthesis. Geological Society of America Bulletin, 88: 857-868.

APPENDIX I
STRATIGRAPHIC SECTIONS EPG-I-VI

LEGEND: MEASURED SECTIONS, APPENDIX I

Mudstone to siltstone; thinly laminated

Siltstone to very fine-grained sandstone; laminated to thinly interbedded

Fine to medium-grained sandstone; thinly bedded to massive

Chert-pebble conglomerate; poorly sorted; matrix-rich; disarticulated fossils and rip-ups

Chert-pebble conglomerate; moderate to well-sorted; inverse to normally graded; imbricated (A-axis parallel to flow)

Chert-pebble to cobble conglomerate; moderate to poorly sorted (rusty conglomerate of fan-delta and coastal assemblage)

Chert-pebble to boulder conglomerate; poorly sorted; internally chaotic; rare inverse grading

ooo

imbricated pebbles

HCS

hummocky cross-stratification

wavy bedding

parallel stratification

ripple crosslamination

trough crossbedding

planar crossbedding

inverse graded bedding

normal graded bedding

sample location

mud rip-ups

trace fossil

fossil

plant fragments

carbonaceous layer

syn-depositional fault

soft-sediment fold

convolute lamination

flame structure

concretion

EPG - 90 -I 380 - 490 m

EPG - 90 - I 600 - 720 m

EPG - 90 - I 720 - 830 m

EPG - 90 -I 830 - 950 m

EPG - 90 - III 414 - 534 m

EPG - 90 - III 534 - 654 m

EPG - 90 - III 654 - 774 m

EPG - 90 - III 1576 - 1696 m

EPG - 90 - IV 0 - 120 m

EPG - 90 - IV 120 - 240 m

EPG - 90 - IV 240 - 360 m

EPG - 90 - IV 360 - 390 m

EPG - 90 - V 0 - 120 m

EPG - 90 - V 360 - 480 m

3 1 of/de 3

PM-1 3 1/2"x4" PHOTOGRAPHIC MICROCOPY TARGET
NBS 1010a ANSI/ISO #2 EQUIVALENT

PRECISIONSM RESOLUTION TARGETS

PIONEERS IN METHYLENE BLUE TESTING SINCE 1974

MicroD

1989 COUNTY ROAD 2, BIRMINGHAM, AL 35207 USA
TEL. 205 485 1927 FAX. 205 485 1927 TLA 940000400

EPG - 90 - V 840 - 960 m

EPG - 90 - V 960 - 1080 m

184

EPG - 90 -VI 0 - 120 m

186

120
100
80
60
40
20
0

.06 2 64 mm

2

ASSEMBLAGE

PRODELTA-SLOPE

Facies

Slope

EPG - 90 - VI 120 - 240 m

EPG - 90 - VI 240 - 360 m

188

EPG - 90 -VI 720 - 840 m

192

EPG - 90 -VI 840 - 960 m

EPG - 90 - VI 960 - 1042.5 m

APPENDIX II
MAXIMUM CLAST SIZE
VERSUS
BED THICKNESS:
STUDY METHODS

Introduction

Sediment gravity flows often play significant roles in the deposition of coarse-grained, gravel-rich deposits. The abundance of pebble- to cobble-sized conglomerates within the Bowser Lake Group prompted a study of the relationship between maximum clast size (MCS) and bedding thickness (BTh). This study is based on the hypothesis, originally proposed by Bluck (1967) and reaffirmed by Nemec and Steel (1984), that a positive linear relationship exists between the MCS (also known as competence) and BTh for conglomerates deposited by sediment gravity flows. The competence is defined as the largest clast size that can be held in suspension above the bed during the life of the flow (Nemec and Steel, 1984). This linear correlation only occurs with sediment gravity flows and does not apply to other environments such as those of fluvial origin (Nemec and Steel, 1984). The study of the two main conglomerate types, the channel conglomerates of the submarine canyon/gully assemblage and the rusty conglomerates of the braid-delta facies, produced poor correlation coefficient values (R) compared to the published studies in Koster and Steel (1984). Because conglomerates of the braid-delta facies display evidence suggesting fluvial deposition (already discussed), poor correlation values would be expected. The channel conglomerates are interpreted to reflect a variety of sediment gravity flow processes in accordance with the downslope change in the internal fabrics of the conglomerates. This results in a relatively poor correlation between MCS and BTh.

Methods

In accordance with Bluck's (1967) method, the MCS represents the mean of the ten largest clast diameters within a bed, measured to the nearest millimetre. Excellent exposure of the conglomerates, in addition to the common occurrence of protruding clasts, provided ideal surfaces (generally 5 to 10 m laterally) to take measurements. Abnormally large clasts, also known as outsized clasts, were not included in the study (Bluck, 1967). Outsized clasts are attributed to rolling or sliding as bedload, during or after cessation of the main flow (Mustard, 1990). Bed thickness (BTh) was measured to the nearest cm in the area where clast diameters were taken. Bed boundaries were discerned primarily on the basis of abrupt clast-size changes, sandstone interbeds, and abrupt changes in grading (primarily in the channel conglomerates).

Statistical Analysis

Data were graphed, analyzed and collated using the program Cricket Graph (version 1.3.2, 1986). The data for the study were separated into two sets, the channel conglomerates and the braid-delta facies. Other conglomerates such as those of the shelf assemblage (interpreted to represent storms) were not included in the study. Figures 6 and 7 (see main text) display best-fit linear regression lines, annotated with corresponding Pearsonian Coefficient values (R), number of samples (n), mean bed thickness (BTh), and mean clast size (MCS).

END

2|5 0|5 |93

FIN