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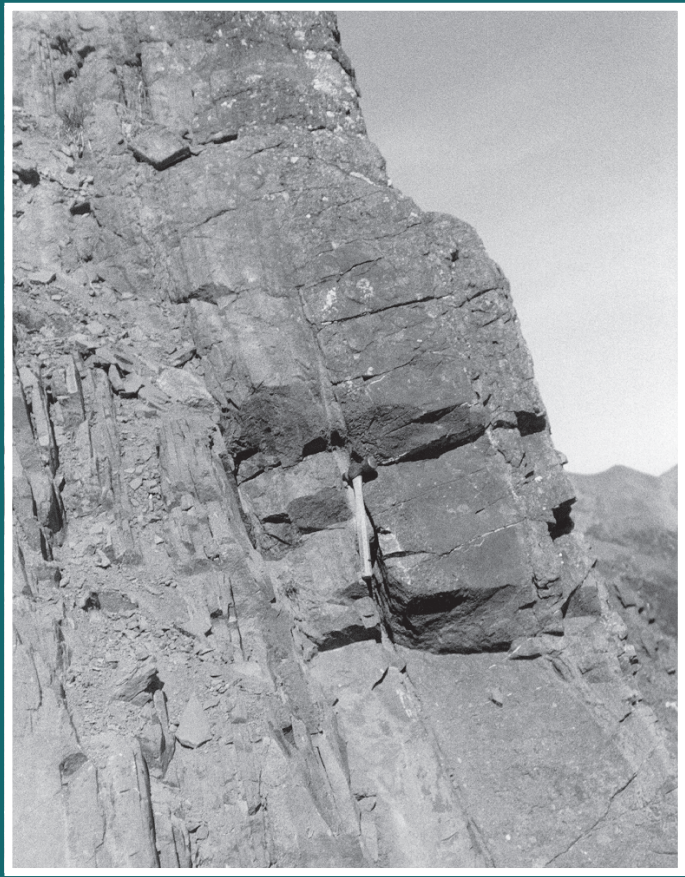
STRATIGRAPHY, DEPOSITIONAL ENVIRONMENT, AND TECTONIC SETTING OF THE UPPER TRIASSIC TO MIDDLE JURASSIC ROCKS OF THE CHILCOTIN RANGES, SOUTHWESTERN BRITISH COLUMBIA

P.J. Umhoefer and H.W. Tipper



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Paul J. Umhoefer and Howard W. Tipper

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Cover illustration

Uppermost Triassic (Late Norian–Rhaetian) sandstones of the Tyaughton Formation on the left unconformably overlain by sandstone and pebble and coarse conglomerate of the Last Creek Formation, Early Jurassic (Late Hettangian), on the right. Exposed on the south side of Castle Pass. Photograph by H.W. Tipper. GSC 1997-52E

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PREFACE

The Upper Triassic to Middle Jurassic rocks of the Chilcotin Ranges in southwestern British Columbia are a unique assemblage of mainly marine sediments that are divisible into formations and members. Macrofossils, mainly ammonoids, and microfossils provide a reliable basis for dating the stratigraphic units. Sedimentary structures and other features provide a basis for determination of the provenance of the sediments and the sedimentary history of the units.

The first geological mapping of this part of the Chilcotin Ranges was by C.E. Cairnes, of the Geological Survey of Canada, in 1939. Since then many studies by other members of the Geological Survey of Canada and members of the British Columbia Geological Survey, as well as by university graduate students, have expanded our knowledge of the stratigraphic units. This report is a summary of the tectonics, stratigraphy, biostratigraphy, paleontology, and sedimentology of these unique sedimentary rocks.

M.D. Everell
Assistant Deputy Minister
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PRÉFACE

Les roches du Trias supérieur au Jurassique moyen des chaînons Chilcotin, dans le sud-ouest de la Colombie-Britannique, constituent un assemblage unique composé essentiellement de roches sédimentaires marines que l'on répartit en formations et en membres. Les macrofossiles (en particulier les ammonoïdes) et les microfossiles offrent une base sûre pour déterminer l'âge des unités stratigraphiques. Les structures sédimentaires et d'autres caractéristiques permettent d'établir la provenance des sédiments et la chronologie stratigraphique des unités.

La première carte géologique des chaînons Chilcotin a été dressée en 1939 par C.E. Cairnes de la Commission géologique du Canada. Depuis lors, de nombreuses études effectuées par d'autres membres de la Commission géologique du Canada et de la Commission géologique de la Colombie-Britannique, ainsi que par des étudiants des cycles supérieurs, ont contribué à élargir les connaissances sur ces unités stratigraphiques. Le présent bulletin constitue une synthèse de la tectonique, de la stratigraphie, de la biostratigraphie, de la paléontologie et de la sédimentologie des roches sédimentaires uniques des chaînons Chilcotin.

M.D. Everell
Sous-ministre adjoint
Secteur des sciences de la Terre

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STRATIGRAPHY, DEPOSITIONAL ENVIRONMENT, AND TECTONIC SETTING OF THE UPPER TRIASSIC TO MIDDLE JURASSIC ROCKS OF THE CHILCOTIN RANGES, SOUTHWESTERN BRITISH COLUMBIA

Abstract

A unique sequence of locally fossiliferous sedimentary rocks is exposed along the northeastern flank of the southern Coast Mountains in the Chilcotin Ranges. The Upper Triassic to Middle Jurassic strata are part of the Cadwallader terrane and have been defined as the Upper Triassic Tyaughton Formation (?Middle and Late Norian age) and Lower to Middle Jurassic Last Creek and Nemaia formations (Late Hettangian to Early Bajocian age). These formations have been further divided into members. Fossils have provided good age constraints. The Tyaughton Formation comprises about 600 m shallow marine limestones and fluvial to shallow marine clastic rocks. The Last Creek Formation disconformably overlies the Tyaughton Formation and comprises at least 400 m of nearshore clastic rocks that grade upward to offshore or slope facies. The newly defined Nemaia Formation is correlative to the Last Creek Formation and comprises about 1.5 km of clastic rocks that grade from locally variable, but dominantly fine grained, Lower Jurassic offshore and slope facies up into Middle Jurassic delta facies.

Résumé

Une séquence unique de roches sédimentaires fossilifères affleurent dans les chaînons Chilcotin, le long du versant nord-est de la partie sud de la chaîne Côtière. Les strates du Trias supérieur au Jurassique moyen, qui font partie du terrane de Cadwallader, ont été associées à la Formation Tyaughton du Trias supérieur (Norien moyen? à tardif) ainsi qu'aux formations de Last Creek et de Nemaia du Jurassique inférieur à moyen (Hettangien tardif-Bajocien précoce), lesquelles ont été subdivisées en membres. Les fossiles ont permis de déterminer les âges de ces roches avec exactitude. La Formation de Tyaughton est composée d'environ 600 mètres de calcaires (sédimentation dans un milieu marin d'eau peu profonde) et de clastites (sédimentation allant d'un milieu fluviatile à un milieu marin d'eau peu profonde). La Formation de Last Creek repose en discordance sur la Formation de Tyaughton et comprend au moins 400 mètres de clastites de faciès littoral, qui passent vers le haut à des faciès de plate-forme continentale ou de talus. La Formation de Nemaia, nouvellement définie, est équivalente à la formation de Last Creek. Elle est constituée de 1,5 kilomètre de clastites qui, de faciès de plate-forme continentale et de talus du Jurassique inférieur (granulométrie variable, mais à dominance fine), passent à des faciès de delta du Jurassique moyen.

SUMMARY

Along the eastern margin of the southern Coast Mountains from Tatlayoko Lake southeastward to Castle Mountain, sedimentary rocks of (?)Middle to Late Norian age (Late Triassic) to Early and early Middle Jurassic age are exposed intermittently. Except for thin, rare tuffaceous beds the sequence is entirely sedimentary. The strata is highly deformed but, because of well preserved macrofossils and microfossils it is possible to reconstruct the stratigraphic succession.

The sedimentary rocks are described and subdivided as three formations and several members. The Upper Triassic part of the sequence is described as the Tyaughton Formation and this is further subdivided into seven lithologically distinct members. The basal red conglomerate member is up to 300 m thick, and composed of limestone and

SOMMAIRE

Des roches sédimentaires, dont l'âge s'échelonne du Norien moyen?-tardif (Trias tardif) au Jurassique précoce-moyen (début), affleurent de façon intermittente dans la partie sud de la chaîne Côtière (le long de la zone orientale), à partir du lac Tatlayoko vers le sud-est jusqu'à la montagne Castle. À l'exception de rares lits tufacés de faible épaisseur, la séquence est entièrement sédimentaire. Les strates sont très déformées, mais il est possible de reconstituer la succession stratigraphique grâce à la présence de macrofossiles et de microfossiles bien conservés.

Les roches sédimentaires sont décrites et subdivisées en trois formations et plusieurs membres. La partie du Trias supérieur de la séquence correspond à la Formation de Tyaughton, laquelle est subdivisée en sept membres lithologiques distincts. Le membre basal composé de conglomérats rouges peut atteindre une épaisseur de 300 mètres. Les fragments de roches calcaires et volcaniques de ces conglomérats passent des gros cailloux aux

volcanic cobble conglomerate grading to pebble conglomerate and to medium grained sandstone and siltstone. The member has a pervasive red to reddish brown coluor and is interpreted to be a series of debris flows possibly formed in an arid climate. Paleocurrent indicators suggest paleoflow to the west and southwest. Overlying the red conglomerate is the massive limestone member which is a 30 to 50 m thick, light grey, thick bedded, bioclastic limestone. This member grades upward into the *Monotis* limestone member, a 0-14 m thick sequence of yellow and grey micrites. This is succeeded by two conglomeratic members, the limestone cobble conglomerate member 0-24 m thick and the lower green clastic member 60-100 m thick. Above these coarse clastic members are the *Cassianella* beds which form a distinct member commonly replete with a diverse fauna of bivalves and ammonites. This member is green-brown- to brown-weathering, fine grained calcareous sandstone to sandy siltstone 90 to 120 m thick. The uppermost member of the formation is called the upper green clastics member consisting of green sandy conglomerate and conglomeratic sandstone 0 to 74 m thick.

Macrofossils and microfossils found in four of the members provide reasonably precise age determinations. The basal red conglomerate member may be as old as Middle Norian and the youngest member, the upper green clastics member, has yielded the youngest known conodonts and ammonites characteristic of the youngest Triassic zone. For the most part the Tyaughton Formation is Late Norian in age; only the basal red conglomerate member is possibly older.

The Jurassic strata near Castle Peak are Early to Middle Jurassic (Late Hettangian to Early Bajocian) and are herein formally named the Last Creek Formation with two members, the Castle Pass and the overlying Little Paradise. The Castle Pass Member is a coarse grained sequence dominantly pebble conglomerate and sandstone at the base becoming finer grained upsection. The Little Paradise Member is dominantly shale and siltstone with rare coarser beds. The Last Creek Formation has abundant, well preserved fossils, particularly in the Castle Pass Member.

The Nemaia Formation is correlative with the Last Creek Formation and is exposed mainly in the Tatlayoko Lake area. It is a sequence of volcanogenic clastic rocks, sandstone, siltstone, shale, and minor tuff and pebble conglomerate. The formation is subdivided into three members, lower, middle, and upper. The lower member is dominantly black shale and siltstone, the middle member is typically siltstone and shale, some tuffaceous siltstone and siliceous beds, and some fine sandstones. The upper member is a coarse grained sandstone and fine pebble conglomerate. Fossils in the Nemaia Formation are sparse but there is sufficient evidence to suggest that the Nemaia and Last Creek formations are essentially correlative.

petits cailloux; s'observent ensuite des grès et des silstones à grain moyen. De couleur rouge pénétrant à brun-rouge, ce membre dériverait d'une série de coulées boueuses vraisemblablement formée sous un climat aride. Les indicateurs de paléocourant semblent révéler des directions de paléoécoulement vers l'ouest et le sud-ouest. Un membre de calcaire bioclastique massif recouvre les conglomérats rouges et atteint de 30 à 50 mètres d'épaisseur; il est de couleur gris clair et présente des lits épais. Il passe vers le haut à un membre calcaire à *Monotis*, qui est constitué d'une séquence de micrites jaunes et grises mesurant jusqu'à 14 mètres d'épaisseur. Ce dernier est suivi d'un membre supérieur de conglomérat calcaire à gros cailloux (jusqu'à 24 mètres d'épaisseur) et d'un membre inférieur de clastites vertes (de 60 à 100 mètres d'épaisseur). Les lits à *Cassianella* sus-jacents constituent un membre distinct; on y observe souvent une faune abondante et diversifiée de bivalves et d'ammonites. Ce membre, dont la couleur d'altération s'échelonne de brun-vert à brun, a une composition allant des grès calcareux à grain fin aux silstones sableux; il mesure de 90 à 120 mètres d'épaisseur. Le membre sommital de la formation, appelé membre de clastites vertes, comprend des conglomérats sableux et des grès conglomératiques atteignant 74 mètres d'épaisseur.

Les macrofossiles et les microfossiles observés dans quatre des membres décrits ci-haut permettent de déterminer leur âge avec une assez grande précision. Le membre basal de conglomérats rouges semble être du Norien moyen; le membre le plus jeune, soit celui de clastites vertes, renferme les conodontes et les ammonites les plus récents qui soient connus, caractéristiques de la partie la plus récente du Trias. La Formation de Tyaughton est en grande partie du Norien tardif; seul le membre basal de conglomérats rouges est peut-être plus ancien.

On a officiellement nommé «Formation de Last Creek» les strates du Jurassique précoce-moyen (Hettangien tardif-Bajocien précoce) observées dans les environs du pic Castle. Cette formation se compose de deux membres, celui de Castle Pass et celui de Little Paradise (sus-jacent). Le Membre de Castle Pass est une séquence à grain grossier prédominée, à la base, par des conglomérats à petits cailloux et des grès, les grains devenant plus fins vers le haut de la coupe. Le Membre de Little Paradise est dominé par des shales et des silstones; s'observent aussi de rares lits à grain plus grossier. La Formation de Last Creek contient de nombreux fossiles bien conservés, en particulier dans le Membre de Castle Pass.

La Formation de Nemaia est équivalente à la Formation de Last Creek; elle affleure principalement dans la région du lac Tatlayoko. Il s'agit d'une séquence comprenant des clastites d'origine volcanique, des grès, des siltstones et des shales, mais aussi accessoirement des tufs et des conglomérats à petits cailloux. La formation se subdivise en trois membres (inférieur, moyen et supérieur). Le membre inférieur est dominé par les shales noirs et les siltstones. Le membre moyen se compose principalement de silstones et de shales et, en moindre abondance, de siltstones tufacés, de lits siliceux ainsi que de grès à grain fin. Le membre supérieur est formé de grès à grain grossier et de conglomérats à très petits cailloux. Dans la Formation de Nemaia, les fossiles sont clairsemés, mais il existe suffisamment d'éléments probants laissant supposer l'existence d'une corrélation importante entre la Formation de Nemaia et celle de Last Creek.

The clastic rocks of the Tyaughton Formation and Last Creek Formation were derived from an inactive magmatic arc and define two distinct petrofacies. The older units in the Tyaughton Formation comprise a mixed plutonic and volcanic derived petrofacies. The second petrofacies is a volcanic derived facies made up of the upper green clastics of the Tyaughton Formation and the overlying Castle Pass Member of the Last Creek Formation. Whereas the older petrofacies is relatively quartz-rich, the younger is quartz poor.

Les clastites des formations de Tyaughton et de Last Creek sont issues d'un arc magmatique inactif; elles définissent deux faciès pétrographiques distincts. Le premier, représenté par les unités les plus anciennes de la Formation de Tyaughton, est d'origine mixte (plutonique et volcanique). Le second, d'origine volcanique, comprend le membre sommital de clastites vertes (Formation de Tyaughton) et le Membre de Castle Pass sus-jacent (Formation de Last Creek). Le faciès plus ancien est relativement riche en quartz alors que le plus récent en renferme peu.

INTRODUCTION

Within the Chilcotin Ranges, along the eastern margin of the southern Coast Mountains from Tatlayoko Lake in the northwest to Castle Peak 150 km to the southeast, a belt 5 to 20 km wide has scattered exposures of (?)Middle to Upper Norian (Upper Triassic) through Lower and early Middle Jurassic sedimentary rocks (Fig. 1). The area is mountainous but generally not rugged, reaching elevations of 2500 to 3000 m. Exposure is good to poor with common extensive scree slopes obscuring the strata. Younger Jurassic and Cretaceous strata and a few Tertiary intrusions further impede the study of the Upper Triassic to Middle Jurassic rocks.

These sequences are totally sedimentary except for thin, rare tuffaceous beds and some coarse volcanogenic strata near Tatlayoko Lake interbedded in the sedimentary sequence. The rocks are unmetamorphosed. Such a section of virtually all sedimentary rocks spanning most of Late Norian to Early Bajocian time is unique in the western Canadian Cordillera; in most areas of the western Cordillera where rocks of this age occur, volcanic rocks are common and prominent in several intervals and many stages are missing. This sedimentary belt is the locus of a multitude of faults and as a result the strata are locally highly deformed, making it difficult to reconstruct the stratigraphic succession in some locations. Despite this, locally common fossils and some distinct lithologies allow accurate correlation of strata and construction of reliable sedimentary sections.

The purpose of this study is to bring together several accounts of work in this belt to better understand the lithostratigraphy, biostratigraphy, sedimentology and tectonic evolution of the Triassic and Jurassic rocks of this unique area. This

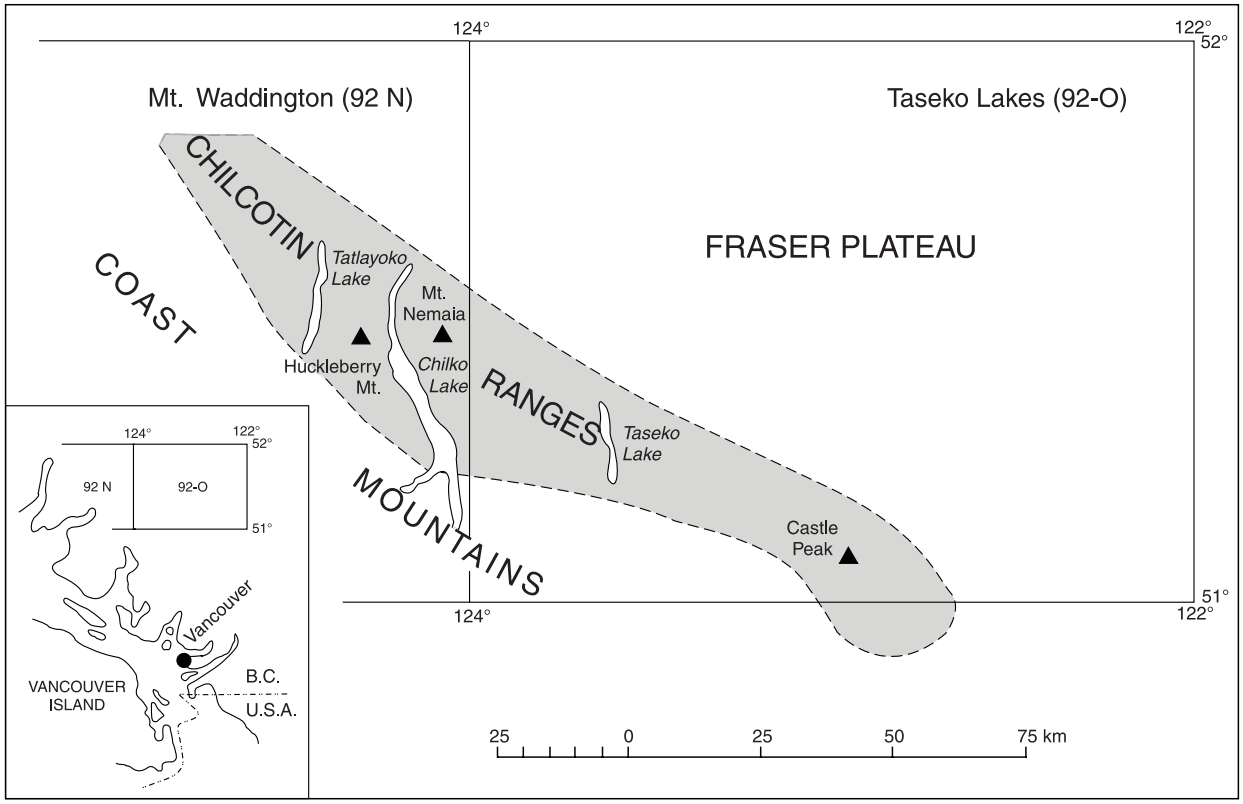


Figure 1. *Locality map of the study area (the shaded area) in the Chilcotin Ranges.*

account will assist in the assessment of the Chilcotin area as a potential hydrocarbon field (C.J. Hickson, R.M. Bustin, and P.A.H. van der Heyden, abstract presented at Cordilleran Geology and Exploration Roundup, Vancouver, British Columbia, 1989).

PREVIOUS AND CURRENT WORK

Reconnaissance and detailed mapping, theses studies, and specific topical studies have been carried out in this region since early in this century but only in recent years has there been sufficiently detailed information obtained to contribute materially to this report. The work falls logically into two separate areas, the northwest end of the belt around Chilko and Tatlayoko lakes and the southeast end around Tyaughton Creek and Castle Peak (Fig. 1). Systematic mapping and study of the northwest area was undertaken in 1967 (Tipper, 1969; Roddick and Tipper, 1985). In the Tyaughton Creek area, the first mapping was by Cairnes (1943). From 1961 to 1964, Tipper mapped the area as a part of the 1:250 000 mapping of the Taseko Lakes map area (Tipper, 1963, 1978) from which several stratigraphical and paleontological reports resulted (Frebold, 1964, 1967; Tozer, 1967; Jeletzky and Tipper, 1968). The British Columbia Geological Survey remapped an area that included the southeastern end of this belt and reports and maps by J.K. Glover, P. Schiarizza and their colleagues were issued (Glover and Schiarizza, 1987; Glover et al., 1987, 1988a, b; Umhoefer et al., 1988; Garver et al., 1989b; and Schiarizza et al., 1989a, b). Umhoefer undertook a stratigraphic and tectonic study of the area from 1985 to 1989 as part of a doctoral study at the University of Washington (Umhoefer, 1989, 1990). P. Schiarizza and co-workers remapped part of the northwest end of the belt in the early 1990s (Schiarizza and Melville, 1995; Schiarizza et al., 1995; Schiarizza, in press). Importantly, maps produced by the British Columbia Geological Survey form the geological base for the present stratigraphic study and should be consulted for locating the units discussed here on a detailed topographic base (Umhoefer et al., 1988; Schiarizza et al., 1993; Schiarizza and Melville, 1995).

This report draws to a greater or lesser extent on many of the aforementioned referenced articles and on unpublished data and fossil reports of the Geological Survey of Canada. Approximately one week was spent by both writers in the Chilko Lake region in 1990 to further clarify ideas.

TECTONIC SETTING OF THE CHILCOTIN RANGES

The Chilcotin Ranges are located northeast of the Coast Mountains and southwest of the Fraser Plateau in northeastern Mount Waddington and southern Taseko Lakes map areas of British Columbia (Fig. 1). The Coast Mountains are underlain by the Coast Plutonic Complex dominated by Jurassic to Tertiary plutonic rocks and remnant Paleozoic and

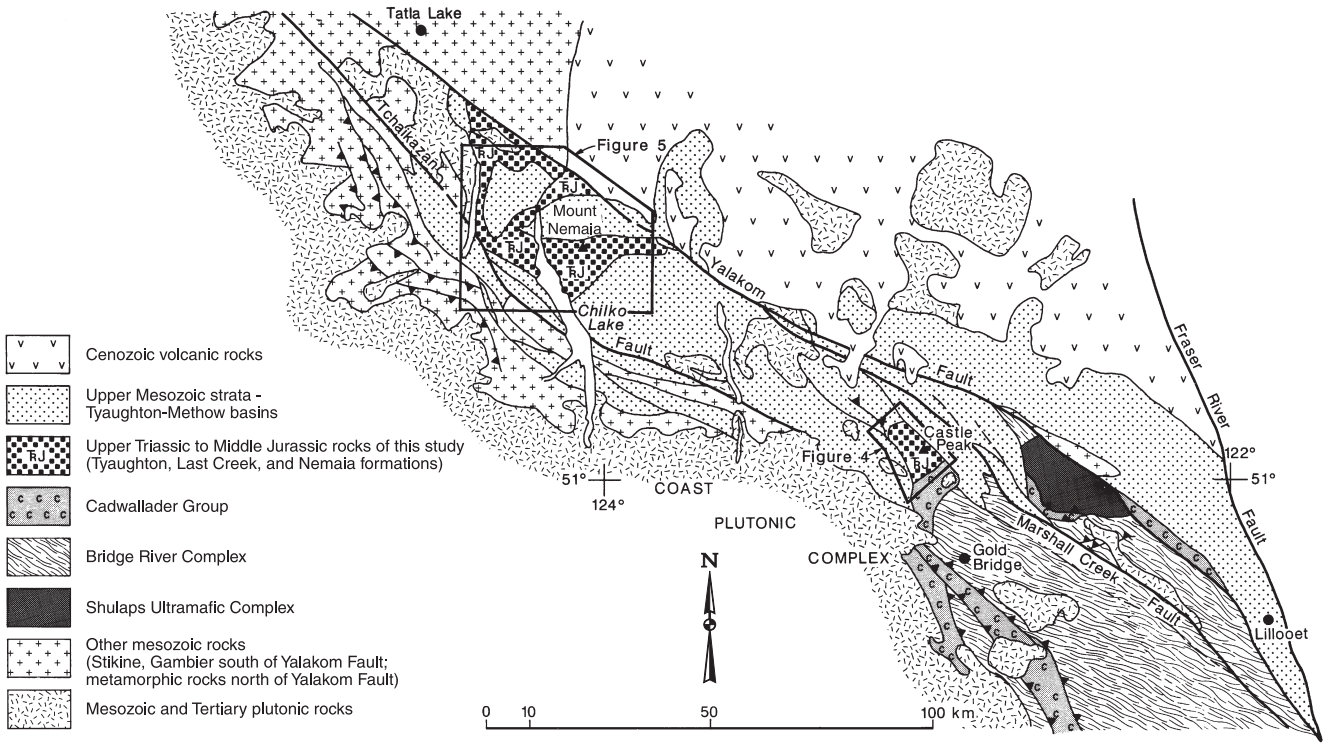


Figure 2. *Generalized geology of the Chilcotin Ranges. Contacts are mostly faults or intrusive contacts. Major faults are labelled and shown with thicker lines. Many thrust faults and strike-slip faults are not shown to simplify map (after Wheeler and McFeely, 1991).*

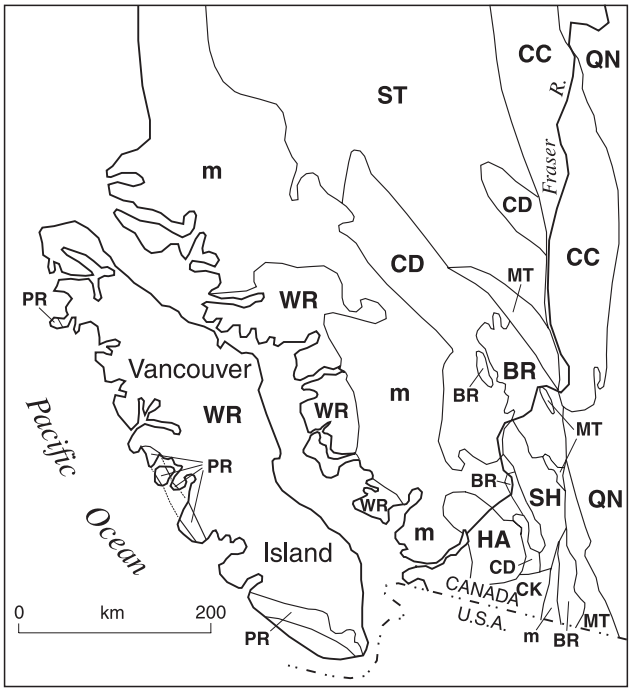
Mesozoic metamorphic rocks (see references in Gabrielse et al., 1991) (Fig. 2). Triassic volcanic and sedimentary rocks exposed along the northeastern margin of the Coast Mountains, immediately southwest of Chilko Lake, have been assigned to the Stikine terrane (Fig. 3) (Rusmore and Woodsworth, 1991a). The Fraser Plateau is underlain by Neogene basalts of the Chilcotin Group (Bevier, 1983), Jurassic and Cretaceous plutons, and lower Mesozoic volcanic and sedimentary rocks (Roddick et al., 1979).

The oldest rocks of the Chilcotin Ranges have been assigned to the Bridge River and Cadwallader terranes, and Shulaps ultramafic complex (Fig. 2, 3). The Bridge River terrane is a Late Paleozoic to Jurassic oceanic assemblage (Fig. 3) (Potter, 1986; Cordey and Schiarizza, 1993). The Shulaps Complex is a structurally disrupted ophiolite of probable Permo-Triassic age (Calon et al., 1990; Schiarizza et al., 1990). The Bridge River terrane and Shulaps Complex both outcrop at the southeastern end of the Chilcotin Ranges, but local lenses of mafic and ultramafic rocks are found along faults of the Yalakom fault system within the Chilcotin Ranges and suggest that these units may underlie other parts of the region.

Throughout most of the Chilcotin Ranges, the oldest rocks are those of the Cadwallader terrane. At the base of the Cadwallader terrane is the Cadwallader Group, including the Pioneer and Hurley formations (Fig. 2, 3). The Pioneer Formation, of latest Carnian and older age, is a sequence of tholeiitic basalts formed in a volcanic arc (Rusmore, 1987). It is overlain by Upper Carnian and/or Lower to Middle Norian tuffaceous sandstones, basalts, and calcareous and volcaniclastic turbidites of the Hurley Formation (Rusmore, 1987). The Hurley Formation indicates the end of Triassic volcanic activity in early Norian time. The Tyaughton Formation, a highly fossiliferous and lithologically distinct sequence of uppermost Triassic clastic and carbonate rocks, is mainly younger than the Hurley Formation but the lower part of the Tyaughton Formation is interpreted to be the nonmarine equivalent of the Hurley Formation (Rusmore, 1987; Umhoefer, 1990). The Lower to Middle Jurassic Last Creek and Nemaia formations overlie the Tyaughton Formation across an erosional unconformity with little or no angular discordance (Umhoefer, 1990); the Jurassic rocks are entirely marine, volcanogenic clastic strata. The Tyaughton, Last Creek, and Nemaia newly defined rock units have been interpreted as comprising the upper part of the Cadwallader terrane, an Upper Triassic to Middle Jurassic sedimentary sequence deposited on the fringe of an inactive, subsiding magmatic arc (Umhoefer, 1990). Current concepts may alter this interpretation somewhat and will be discussed later.

Rusmore et al. (1988) correlated the Cadwallader with the Stikine terrane and the Bridge River with the Cache Creek terrane. They included all of these terranes in the Intermontane superterrane which accreted to North America in the Early to Middle Jurassic (Monger et al., 1982). Cordey and Schiarizza (1993) disputed the correlation of the Bridge River and Cache Creek terranes. They pointed out that the main Cache Creek terrane has no known Jurassic rocks and the western Cache Creek terrane has Early or Middle Jurassic rocks (based on radiolaria from tuffaceous argillite), whereas the Bridge River rocks contain Early and Middle Jurassic radiolaria in ribbon chert. In addition, the Bridge River rocks occur structurally beneath and west of the Cadwallader terrane and therefore can not easily be restored against the Cache Creek terrane, which lies east of the Cadwallader terrane (Cordey and Schiarizza, 1993).

Unconformably overlying the Last Creek and Nemaia formations of the Cadwallader terrane are strata of the Tyaughton basin (Fig. 2). The basal unit of the Tyaughton basin is the Middle Jurassic to Lower Cretaceous Relay Mountain Group, a sequence of clastic rocks deposited primarily in a shallow marine setting (Jeletzky and Tipper, 1968; Tipper, 1969; Umhoefer, 1989). The upper Lower Cretaceous Taylor Creek Group, overlying the Relay Mountain Group, is a synorogenic, marine clastic unit that was derived from the western and eastern margins of the basin and within it (Garver, 1989, 1992). A major disconformity to angular unconformity separates the Taylor Creek Group from overlying lower Upper Cretaceous nonmarine clastic and volcanic rocks (Garver, 1992).



CC	*Cache Creek*	**MT**	*Methow*
QN	*Quesnellia*	**SH**	*Shuksan*
ST	*Stikinia*	**WR**	*Wrangellia*
BR	*Bridge River*	**PR**	*Pacific Rim*
CD	*Cadwallader*	**m**	*Undivided metamorphic rocks*
CK	*Chilliwack*		
HA	*Harrison*		

Figure 3. *Terranes of southwestern British Columbia (after Wheeler and McFeely, 1991).*

The thick, mid-Cretaceous clastic and volcanic units in the Tyaughton and Methow basins provide strong evidence for major contractional deformation in the Albian to Cenomanian, both adjacent to, and within the basins (Garver et al., 1988; McGroder, 1989; Garver, 1989, 1992). A fold-and-thrust belt of both east and west vergence shortened both basins from middle Albian to Santonian time (approximately 100-85 Ma, Schiarizza et al., 1990) resulting in 50 km of shortening in the Methow basin (McGroder, 1989) and possibly a few tens of kilometres of shortening in the southeastern Tyaughton basin (Umhoefer, 1989). Just west of the Tyaughton basin is a northeast-vergent thrust belt which also was active in the Late Cretaceous (Rusmore and Woodsworth, 1988, 1991b, 1994). These thrust belts are coeval with major west-vergent thrust faulting on the western side of the Coast Plutonic Complex from northwestern Washington to southeastern Alaska (e.g. Rubin et al., 1990; McGroder, 1991; Journeay and Friedman, 1993). All of the contractional structures throughout the length and width of the Coast Plutonic Complex were probably formed during the culmination of the accretion of the Insular to the Intermontane superterrane in mid-Cretaceous time (Monger et al., 1982), which may have been the end of a long-lived Mesozoic orogeny (van der Heyden, 1992). The rocks of the Chilcotin Ranges lie on the eastern margin of that deformation belt.

Small Late Cretaceous to early Tertiary plutons intruding the mid-Cretaceous thrust faults and older rock units are distributed throughout the Tyaughton basin. These igneous rocks are spatially associated, and broadly coeval, with the Yalakom fault system, a major dextral strike-slip fault system which cuts all Mesozoic rock units in the Chilcotin Ranges (Fig. 2) (Kleinspehn, 1985; Umhoefer and Schiarizza, 1993).

The youngest rock unit in the Chilcotin Ranges is the Chilcotin Group, a widespread series of plateau basalts of Neogene age (Bevier, 1983). The Chilcotin basalts form gently northeast-dipping sequences on the summits and flanks of some of the highest peaks in the Chilcotin Ranges. The basalts overlie the Yalakom Fault in many places without structural dislocation.

STRATIGRAPHY

General statement

The Upper Triassic and Lower to lower Middle Jurassic rocks of the Chilcotin Ranges have been deformed at least twice, and are exposed in fault-bounded domains of 100 m to a few kilometres. As a result of this structural complexity, no one area exposes the entire Triassic to Middle Jurassic sequence. The stratigraphic succession presented herein is a compilation of a number of measured sections in two principal areas.

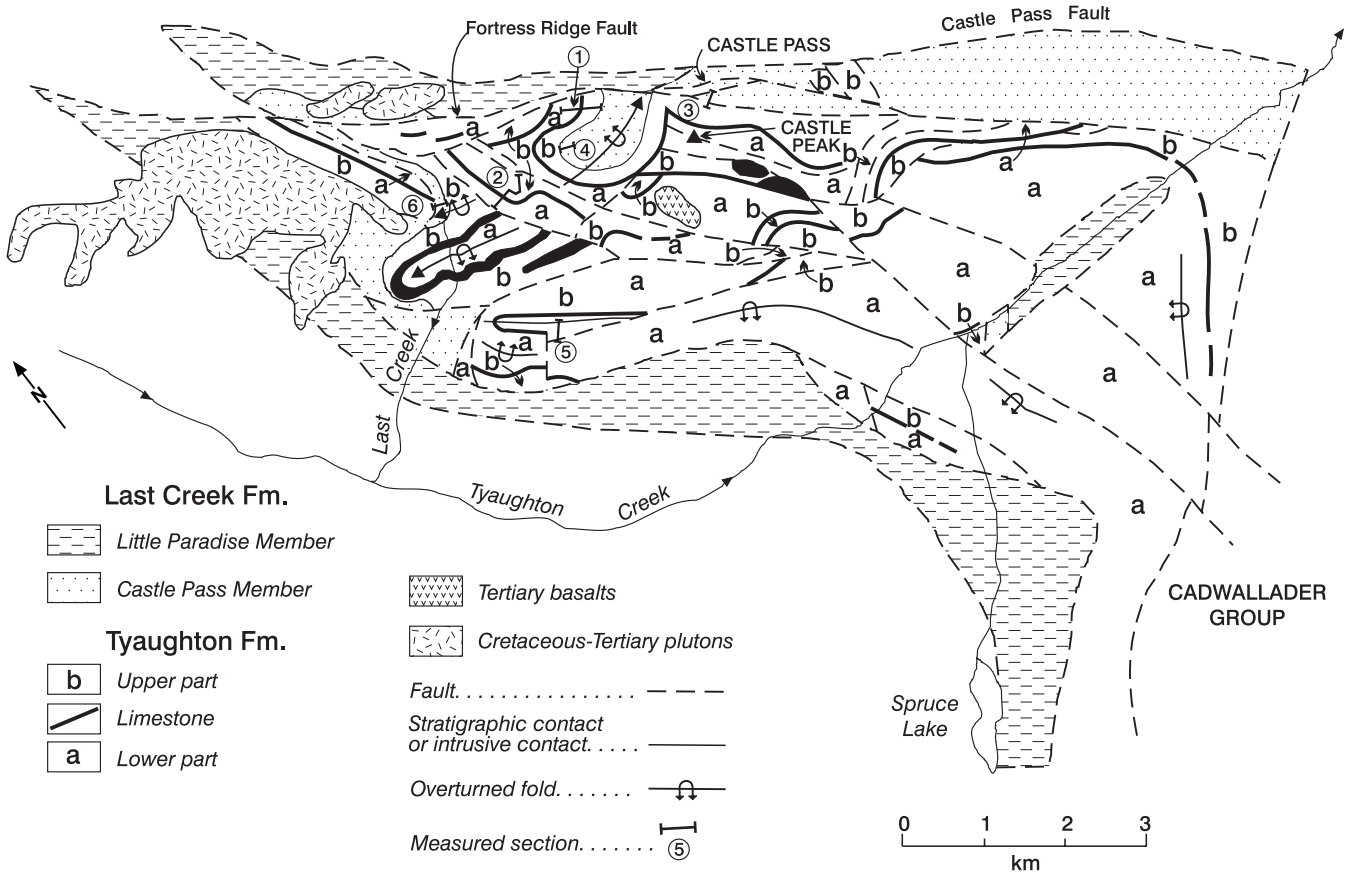


Figure 4. *Geological map of Tyaughton and Last Creek formations in the Castle Peak area. Modified from Umhoefer et al. (1988).*

The Triassic Tyaughton Formation and Jurassic Last Creek Formation are well exposed near, and up to 5 km to the west of Castle Peak in the southern Taseko Lakes map area (Fig. 4). The Jurassic Nemaia Formation is well exposed in two areas on either side of Chilko Lake in the northeastern part of the Mount Waddington map area (Fig. 5). Despite the structural complexity, all three rock units are represented by sections which depict the major part of each unit.

Tipper mapped these units at 1:250 000 scale in the course of mapping the Taseko Lakes map area (Tipper, 1963, 1978) and Mount Waddington map area (Tipper, 1969). The Tyaughton Formation and basal Castle Pass Member of the Last Creek Formation were studied in the greatest detail. Umhoefer (1989) measured three sections in the Tyaughton Formation and a member in the lower part of the Last Creek Formation at a scale of about 1:150. Many other sections were observed and generally described over the course of mapping in these units for almost three months in the summers of 1985, 1986, and 1987.

The upper member of the Last Creek Formation and the entire Nemaia Formation were observed in considerably less detail. Our knowledge of the upper Last Creek Formation is from mapping of the Castle Peak area by Umhoefer at 1:10 000 scale (Umhoefer et al., 1988), and frequent visits to the unit by Tipper since the 1960s to collect ammonites in the fossil-rich shale beds.

The Nemaia Formation was studied in the field in a preliminary fashion for about ten days in 1990. Two long descriptive traverses were completed through the upper part of the Nemaia Formation, as well as observations at a number of other short sequences; a nearly 500 m thick section of the upper member was measured.

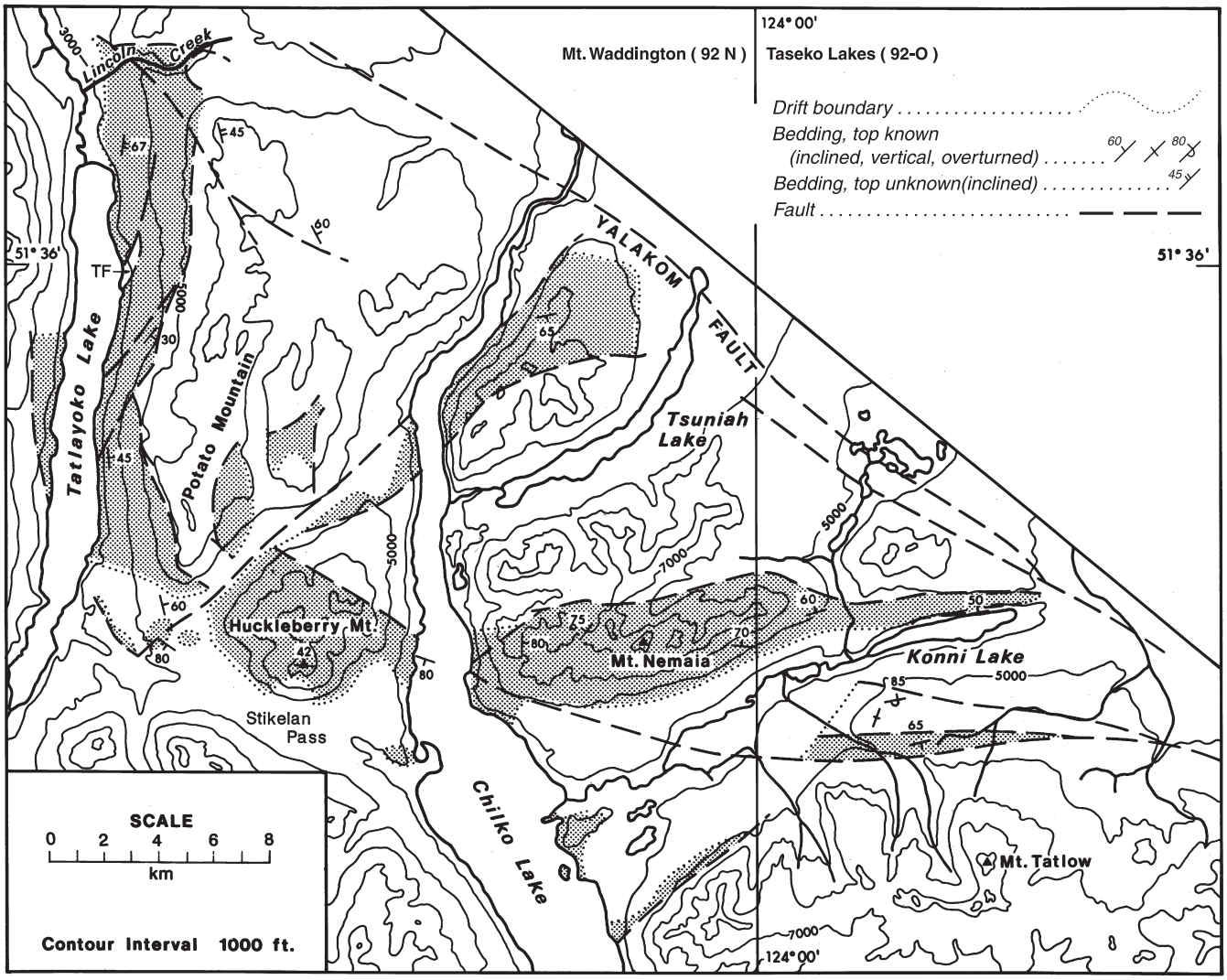


Figure 5. Distribution of Nemaia (grey shading) in Chilko Lake area. Tyaughton Formation along the northeast side of Tatlayoko Lake is the patch of Tyaughton Formation discussed in the text. See Schiarizza and Melville (1995) and Schiarizza (in press) for more details.

TYAUGHTON FORMATION	MEMBERS	CHARACTERISTIC LITHOLOGY	THICKNESS	AGE
	Upper green clastics	green sandy conglomerate, medium-grained sandstone	0-74 m	Late Norian, Crickmayi zone
	Cassianella beds	green-brown to brown calcareous sandstone, sandy siltstone	90-120 m	Late Norian, Amoenum zone
	Lower green clastics	green, medium- to coarse-grained lithic sandstone	60-100 m	Late Norian
	Limestone cobble conglomerate	grey limestone cobble and pebble conglomerate	0-24 m	Late Norian
	Monotis limestone	yellow and grey thin-bedded micrite	0-14 m	Late Norian, Cordilleranus zone
	Massive limestone	light to medium grey thick-bedded, bioclastic limestone	30-50 m	(?)Middle to (?)Late Norian
	Basal red conglomerate	cobble conglomerate, pebble conglomerate, sandstone; red, purple, maroon, red-brown	³300 m (base not exposed)	(?)Middle to (?)Late Norian

Figure 6. *Members of the Tyaughton Formation.*

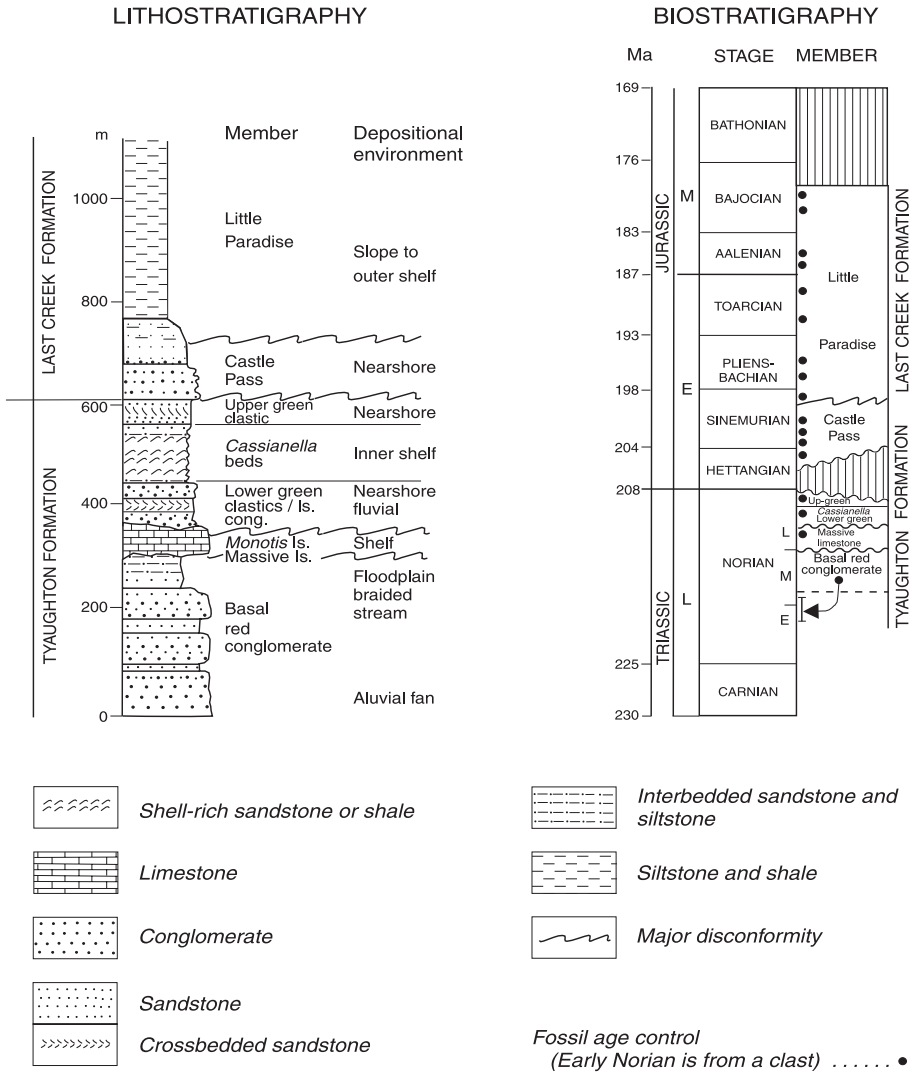


Figure 7. *Biostratigraphic and lithostratigraphic columns of the Tyaughton Formation and Last Creek Formation, Tyaughton Creek area (stages in biostratigraphy column are abbreviated; Up = upper; ls., Ls. = limestone; cong. = conglomerate).*

Tyaughton Formation (new)

Since the initial study of the Tyaughton Group by Cairnes (1943), much biostratigraphic information has been obtained (Tozer, 1967), the area of outcrop has been defined (Tipper, 1963, 1978; Umhoefer et al., 1988; Schiarizza et al., 1989a, b) and division into smaller lithological units has been achieved. The area of outcrop is only about 60 km^2 and the thickness of the sequence is about 600 m. It is not known reliably that it outcrops anywhere but in the type area, except for the small patches in the Mount Waddington map area. For these reasons it is proposed that the unit be reduced to a formation and the subdivisions mentioned above be defined as members. The Tyaughton Formation and its members are shown in Figure 6.

In 1963 E.T. Tozer studied the Tyaughton strata and its fauna. He published two measured sections (Tozer, 1967) and the faunal sequence. He later proposed a revised ammonoid zonal scheme for the time in which the Tyaughton Formation was deposited (Tozer, 1979). His interpretations are essentially unchanged in our definition of the formation.

Definition of the Tyaughton Formation

Castle Peak area

The Tyaughton Formation is an Upper Triassic ((?)Middle to Upper Norian) sequence of distinctive clastic rocks and limestones best exposed on the ridges west and south of Castle Peak in the southern Taseko Lakes map area (Fig. 2, 4) (Umhoefer et al., 1988). The formation is little affected by metamorphism, except near intrusive bodies. Sandstones show only diagenetic alteration of grains by clays and grain-to-grain sutures. A few samples in the lowest unit of the Tyaughton Formation contain zeolites and one has prehnite. Conodont alteration indices (CAI = 1-3 or <50° to 175°C) confirm minimal burial of the unit (M.J. Orchard, GSC, pers. comm., 1988). Despite the lack of metamorphism, the Tyaughton Formation has undergone multiple deformations into broadly to tightly folded blocks bounded by moderate to high angle faults (Fig. 4). No single locality preserves the entire stratigraphic section of the Tyaughton Formation, but three individual sections can be internally correlated so that a minimum unit thickness is reliably estimated at about 600 m (Fig. 7). Most of the formation is exposed just north of Castle Peak in a section greater than 440 m thick (Fig. 8) but the lower redbeds are only 200 m thick and the topmost upper green clastics member is missing due to erosion just after Triassic time (much of this section is in Fig. 9). A 320 m thick

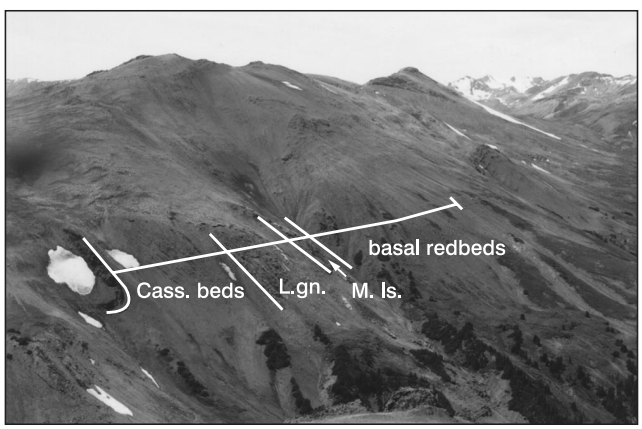


Figure 8. *View northwest towards the slope ~1 km northwest of Castle Pass in upper Paradise Creek valley. The slope contains one of the type sections of the Tyaughton Formation (#1 in Fig. 4), which here includes the basal red conglomerate member (basal redbeds), massive limestone member,* Monotis *limestone member (M.ls), limestone cobble conglomerate member, lower green clastics member (L.gn), and the* Cassianella *beds member (Cass. beds). The dark hooked-shaped outcrop between two snow patches on the left is the base of the Last Creek Formation, which here overlies the* Cassianella *beds with the upper green clastics member of the Tyaughton Formation missing. Photograph by P.J. Umhoefer. GSC 1997-53A*

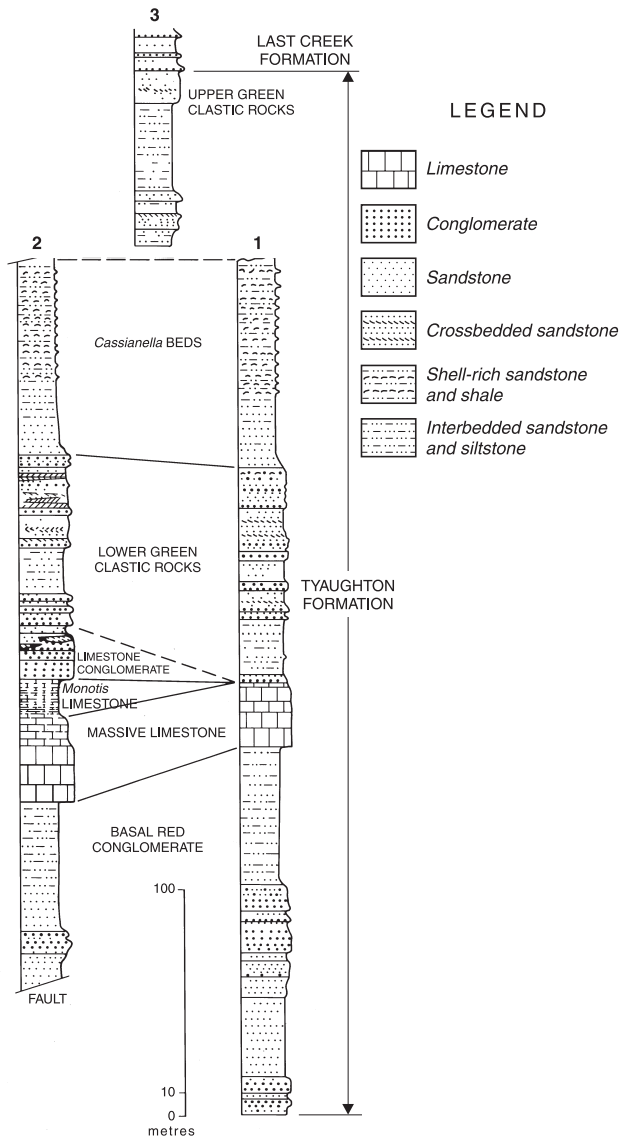


Figure 9. *Schematic sections of Tyaughton Formation. Section locations shown as circled numbers on Figure 4.*

section nicely exposes the middle part of the unit on a ridge 2.5 km northwest of Castle Peak (#2 in Fig. 4, 9, 10). The upper part of the section is well exposed in a number of locations, including the ridge northeast of Castle Pass, where it is overturned with the overlying Last Creek Formation (#3 in Fig. 4; 11, see below).

The stratigraphic base of the Tyaughton Formation is not exposed; everywhere the base of the basal red conglomerate member is a fault or the core of a fold. The redbeds are probably the nonmarine equivalent of the uppermost Hurley - Formation of the Cadwallader Group. The Tyaughton Formation above the basal redbeds is known to be entirely younger than the Hurley Formation exposed nearby and thus may have been deposited over it. Support for deposition of the Tyaughton Formation and Hurley Formation in the same basin (also suggested by Cairnes, 1943; Rusmore, 1987; Rusmore et al., 1988) includes (1) the Middle Norian age for both the uppermost Hurley Formation (Rusmore, 1987) and the basal red conglomerate member of the Tyaughton Formation (see discussion below); (2) the overall similarity in provenance of sandstones and conglomerates of the two units, particularly the lithologically similar limestone clasts which yield some of the same species of conodonts, and (3) juxtaposition of the Hurley Formation and Tyaughton Formation across a thrust fault (Rusmore, 1987; Garver et al., 1989a), which is similar to faults within the Tyaughton Formation with only a few hundred metres of stratigraphic separation. If the Tyaughton Formation above the redbeds was deposited directly on the Cadwallader Group, an erosional unconformity probably separated the two units because the basal rocks of the Tyaughton Formation are nonmarine alluvial-fluvial deposits whereas the uppermost rocks of the Hurley Formation are submarine turbidites.

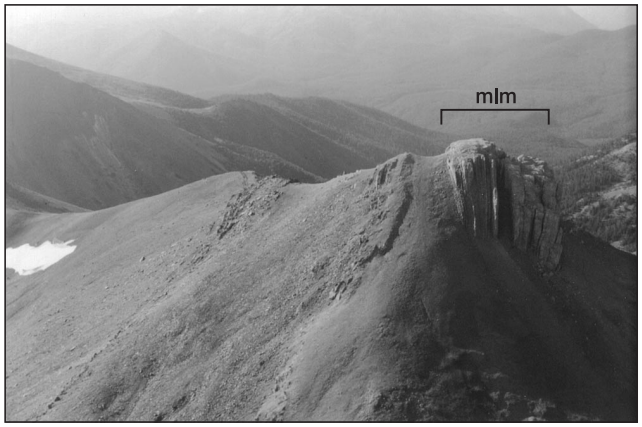


Figure 10. *Looking south at one of the type sections (#2 in Fig. 4) of the middle part of Tyaughton Formation. The ridge is 2.5 km northwest of Castle Peak (Fig. 4) and includes the uppermost basal red conglomerate member (extreme right) up to the* Cassianella *beds member (to the left). The prominent thick-bedded unit on the right (mlm) and making up the highest knob on the ridge is the Massive limestone member, which is 40 m thick. Photograph by P.J. Umhoefer. GSC 1997-53B*

The Tyaughton Formation is disconformably overlain by the Lower to Middle Jurassic Last Creek Formation (Umhoefer et al., 1988). In some locations the upper green clastics member of the Tyaughton Formation and/or basal conglomerate beds of the Last Creek Formation are missing (Fig. 7).

The Tyaughton Formation was divided into five informal map units in earlier mapping (Tozer, 1967) and also on more recent maps (Umhoefer et al., 1988; Garver et al., 1989a). In this report the formation has been subdivided into seven members. The basal red conglomerate member is at least 300 m thick and is best exposed on the slope 1 to 2 km northwest of Castle Peak (#1, Fig. 4), in the core of the anticline on the west side of the Tyaughton Formation outcrop area, and on the top and west side of the ridge due east of Last Creek, though the unit outcrops over much of the map area (Fig. 4).

The massive limestone member is a medium- to thick-bedded bioclastic limestone. In a few places it grades up to a thin bedded *Monotis*-bearing micrite, the *Monotis* limestone member, best exposed on a ridge 2.5 km northwest of Castle Peak (Fig. 4) (Tozer, 1967). The two members vary from 25 to 55 m in thickness, typically 30 to 40 m thick. The massive limestone member overlies the basal red conglomerate member across a disconformity with no evidence of erosion, and underlies the limestone conglomerate and lower green clastics members across an erosional unconformity which cuts out the *Monotis* limestone and the top of the massive limestone member in many places.

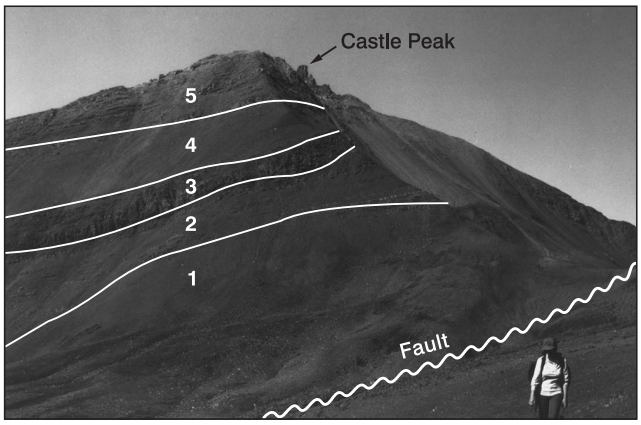


Figure 11. *Overturned section of Last Creek and Tyaughton formations on northeast side of Castle Peak. Castle Pass is on the right. From bottom to top (younger to older): (1) shale-siltstone of Last Creek Formation, Lower Sinemurian (*Coroniceras *beds mainly); (2) pebble conglomerate, grit, sandstone of basal Last Creek Formation; upper Hettangian (Canadensis zone in part and slightly lower); (3) green grit, pebbly sandstone, sandstone of the upper green clastics member of the Tyaughton Formation, Late Norian (Rhaetian); (4)* Cassianella *beds are calcareous siltstone and sandstone of Late Norian age (Amoenum Zone); (5) limestone cobble conglomerate member and lower green clastics member. Photograph by H.W. Tipper. GSC 1997-52A*

The limestone cobble conglomerate member and lower green clastics member grade into the overlying *Cassianella* beds member across a transition section a few metres thick. The lower green clastics member is 60 to 100 m in thickness; the basal limestone-conglomerate member varies from 0 to 24 m. The *Cassianella* beds unit is a distinct sequence of recessive, fine grained sandstone interbedded with prominent ribs of sandy, shell-rich calc-arenite. The unit is 90 to 120 m thick and grades up into the upper green clastics member, the top unit of the Tyaughton Formation.

Chilko Lake area

Beyond the type area, the Tyaughton Formation is found in one small patch in the Chilcotin Ranges, about 120 km northwest of the type area near Castle Peak. In the western Taseko Lakes map area, about 3 km southeast of Konni Lake and possibly extending east to Elkins Creek, discontinuous exposures of limestone and clastic rocks were correlated with the massive limestone and basal conglomerate members of the Tyaughton Formation (Tipper, 1978). However these exposures are part of a belt of Hurley Formation (Riddell et al., 1993) because one sample of limestone yielded Middle Norian conodonts (M.J. Orchard, GSC Internal Report, OF-1993-41) which are only compatible with the Hurley Formation. The lithologies of this belt also match the Hurley Formation.

Along the shore of Tatlayoko Lake, about 3 km south of the northeast corner of the lake (TF on Fig. 5), a 200 m thick sequence of limestone and mainly coarse grained clastic rocks are correlated with the *Cassianella* beds and upper green clastics member of the Tyaughton Formation. From the base, along the lake, upsection to the east, the section includes about 40 m of lithic sandstone, laminated fine grained sandstone, and interbedded siltstone, with uncommon pelecypod-rich, calcareous sandstone to sandy limestone, and minor quartz and volcanic-granular conglomerate, succeeded by about 70 m of covered interval, and then about 100 m of green to grey, granular to medium-grained lithic sandstone, with a thin, dark grey micrite at the base and 3 m of pelecypod-rich sandstone near the top. Pelecypods from this section include *Neomegalodus* and *Cassianella* species (Tipper, 1969), similar to better preserved specimens from the Tyaughton Formation near Castle Peak. The upper contact of this section with the nearby Nemaia Formation is not exposed.

Lithostratigraphy and depositional environment of the Tyaughton Formation

Basal red conglomerate member

This member is up to 300 m thick and displays a general fining-upward from cobble conglomerate to interbedded pebble conglomerate and sandstone, to fine grained sandstone and siltstone in the top 40 to 65 m. Facies contacts are gradational.

The lower pebble- to cobble-conglomerate facies consists of amalgamated red to green conglomerate beds (typically 1-10 m thick) with minor pebbly sandstone interbeds (Fig. 9). The conglomerate of this facies is generally clast-supported, poorly sorted, ungraded to crudely graded, and unstratified to poorly stratified (Fig. 12, 13). Within the limits of exposure, the conglomerate units are generally tabular in cross-section, and amalgamated beds range up to 35 m in cumulative thickness; one lenticular bed was observed to change from 5 m to 1 m over 25 m along bedding. Imbrication of clasts is crudely displayed in a few outcrops. Subangular to subrounded clasts (in all facies of the basal red conglomerate) are dominantly either volcanic (felsic to mafic) or grey and pink limestone, with minor plutonic types (mainly granitoid). Maximum clast size (average of ten largest clasts) and the size of the single largest clast at four scattered localities are 268 mm and 670 mm (Fig. 13, 14), 184 mm and 265 mm (Fig. 12), 112 mm and 145 mm, and 80 mm and 140 mm respectively.

The most common facies in the basal red conglomerate member comprises interbedded pebble conglomerate and pebbly to medium grained sandstone in the middle part of the map unit. The rocks of this facies are typically red, but some exposures are grey-green. Bedding is dominantly

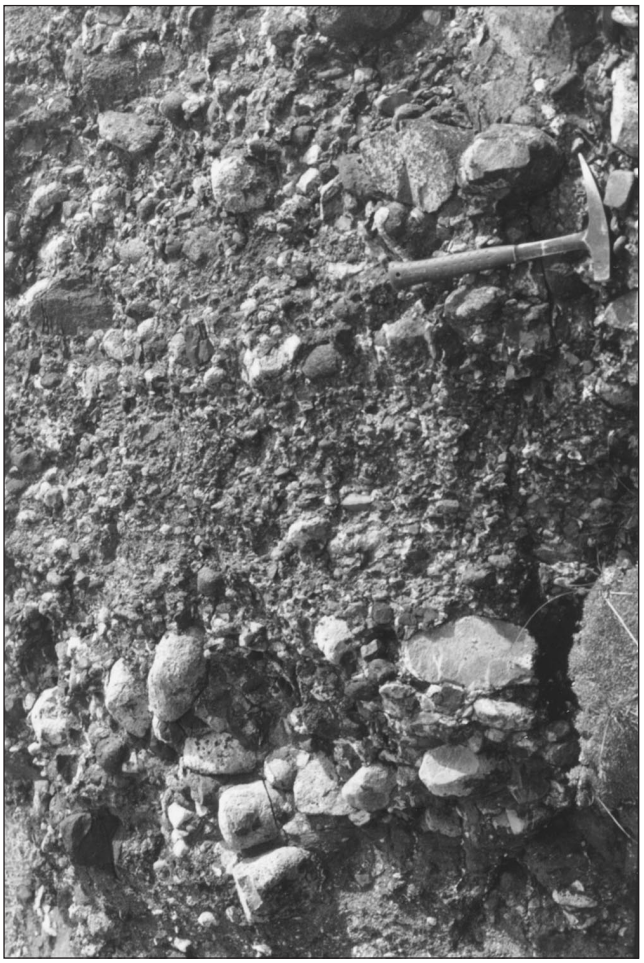


Figure 12. *Photograph of the alluvial facies, basal red conglomerate member of the Tyaughton Formation. Part of a poorly sorted conglomerate. See text for discussion of the clast with the long axis oriented vertically in the lower part of the photograph. Bedding is parallel to the hammer handle. Photograph by P.J. Umhoefer. GSC 1997-53C*

planar-horizontal to planar-inclined; beds are typically a few decimetres in thickness (Fig. 15), ranging up to 1-2 m. Some finer grained beds show parallel laminae but most beds are massive to crudely stratified. Graded conglomerate beds and fining-upward conglomerate to sandstone units across a few metres are common. Conglomerate commonly grades into sandstone laterally. Crossbedded units include tabular sets (Fig. 16), especially in the thicker units, and trough sets within otherwise planar, thin-bedded, coarse sandstone units (Fig. 17). Pebble conglomerate beds are moderately sorted, with most clasts less than 10 cm in maximum size; clasts are commonly oriented parallel to bedding; well-developed imbrication is uncommon. South of Tyaughton Creek a variation of this facies includes at two localities, thin-bedded brown micrites containing possible fresh water bivalves.

The upper 40-65 m of the member are poorly exposed, purple-red to red, fine- to medium-grained sandstone and sandy siltstone (Fig. 17). Most outcrops display thin, planar bedding and horizontal laminae; carbonate concretions are present but uncommon.

The basal red conglomerate member is interpreted to be an alluvial fan to braided stream sequence reflecting initial coarse-clastic input due to uplift of a nearby volcanic, limestone, and plutonic source area, followed by a general diminishing of source-area relief and retreat of the fault scarp and alluvial fan. Characteristics of the basal red conglomerate, similar to well documented modern and ancient alluvial fans (e.g. Nilson, 1982), are the rapid decrease in average and maximum grain size upsection (interpreted to be down fan),

Figure 13. *Tyaughton Formation, basal red conglomerate member. Hammer is parallel to bedding. This photo is a close up of lower right of Figure 14. Photograph by P.J. Umhoefer. GSC 1997-53D*

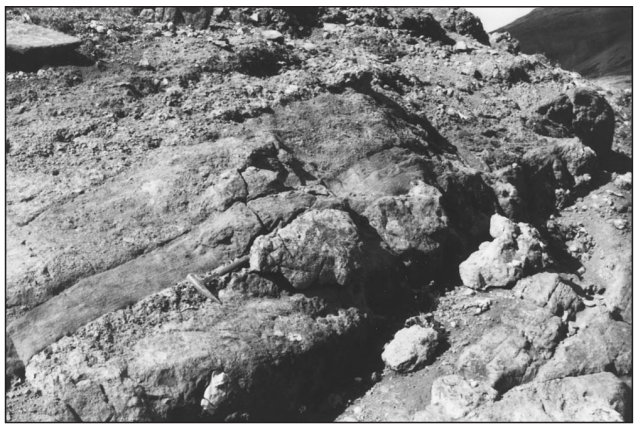


Figure 15. *Photograph of the planar-bedded sandstone and conglomerate, basal red conglomerate member of the Tyaughton Formation. The bed under the hammer is graded from pebble conglomerate to pebbly sandstone. Beds are overturned and top is to the lower right. Photograph by P.J. Umhoefer. GSC 1997-53F*

Figure 14. *Photograph of the alluvial facies, basal red conglomerate member of the Tyaughton Formation. Amalgamated thick to massive beds of cobble to boulder conglomerate. Photograph by P.J. Umhoefer. GSC 1997-53E*

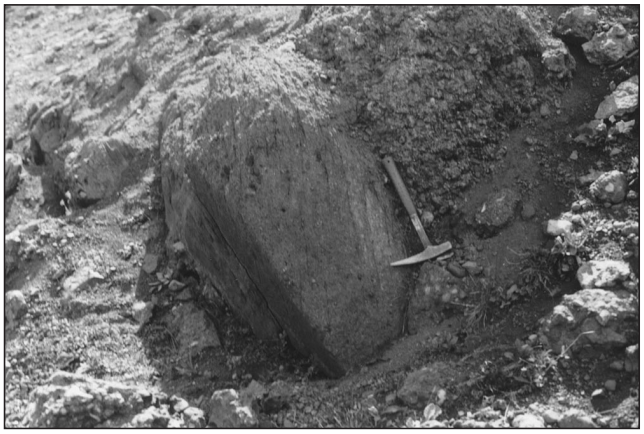


Figure 16. *Photograph of overturned bed of graded conglomerate and crossbedded pebbly sandstone, crossbedded and parallel-laminated sandstone, and pebbly sandstone; basal red conglomerate member of the Tyaughton Formation. Photograph by P.J. Umhoefer. GSC 1997-53G*

Figure 17.

Photograph of crossbedded and parallel-laminated sandstone and pebbly sandstone, basal red conglomerate member of the Tyaughton Formation. Tops are to the right. Stick is 1.4 m. Photograph by P.J. Umhoefer. GSC 1997-53H

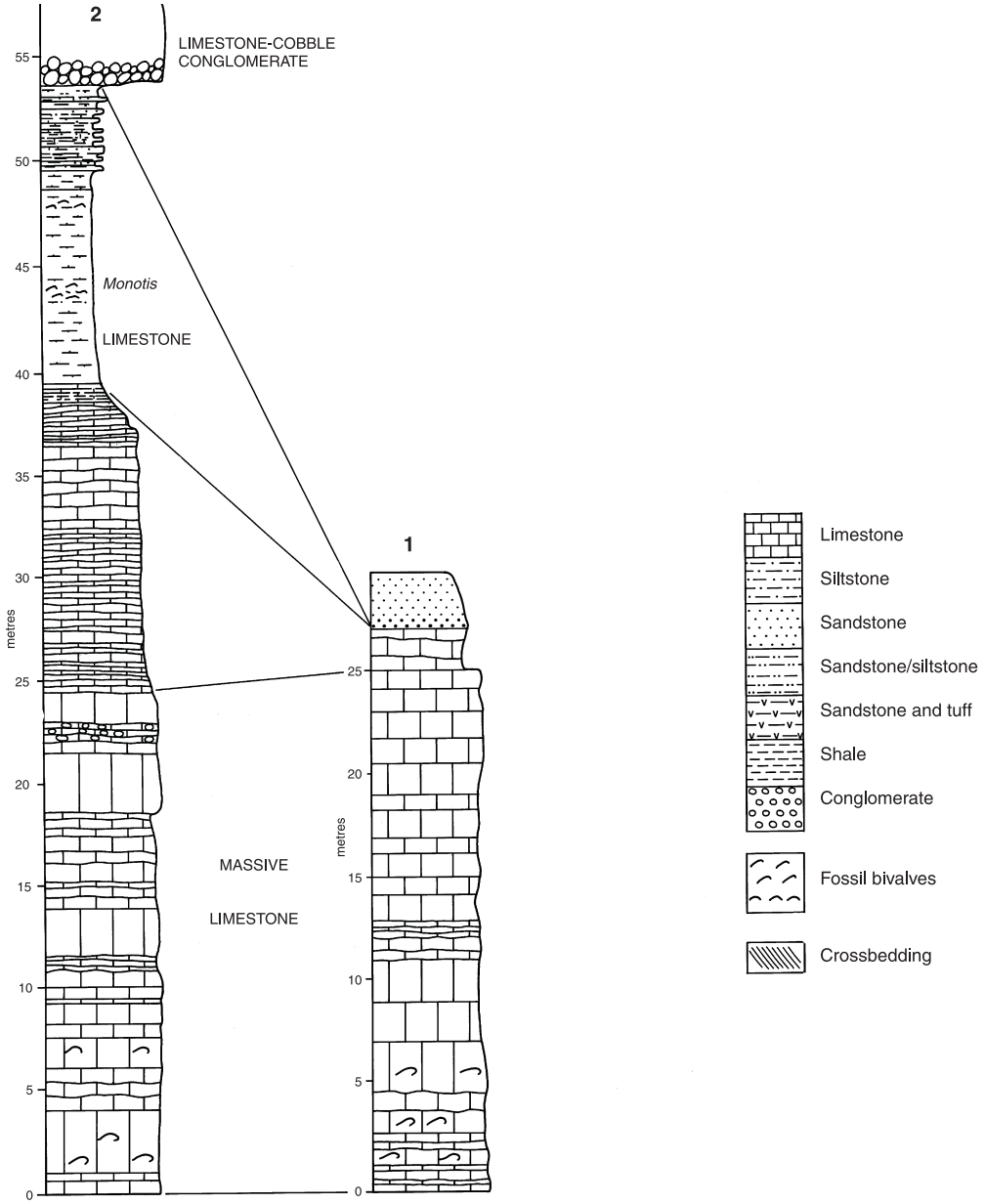


Figure 18. *Measured sections of the massive limestone and* Monotis *limestone members of the Tyaughton Formation. Section locations are shown as circled numbers on Figure 4.*

compositionally immature sedimentary rocks, oxidized sedimentary rocks (red and yellow-red), dominance of planar stratification and large-scale cross-strata, dominance of both coarse unsorted deposits (debris flow) and moderately sorted deposits (streamflow and sheetflood), and the lack of fossils and organic matter.

The basal cobble conglomerate facies is interpreted to be a series of debris flows and possibly hyperconcentrated flood deposits (as in Smith, 1986), typical of the proximal- to mid-fan region on alluvial fans (e.g. Nilson, 1982; Nemec and Steel, 1984). The overlying pebble conglomerate and sandstone facies is interpreted to be proximal, braided stream deposits, and the upper, fine grained facies as sandy, braided stream and floodplain deposits, with possible minor paleosols. The few paleocurrent measurements indicate paleoflow to the west and southwest after structural correction by simple rotation of bedding to horizontal.

A number of aspects of the Tyaughton Formation redbeds suggest that the alluvial fan might have formed in an arid climate. The common red coloration due to iron oxide diagenesis has been argued as being indicative of arid conditions (Walker, 1967; Van Houten, 1968). Abundant limestone clasts in the conglomerates are most readily preserved in an arid climate. Alluvial fans along the Gulf of Aqaba in the Middle East (Hayward, 1985) may be a good analogy to the basal Tyaughton sequence. At the margin of the Gulf of Aqaba, 2-4 km wide coastal fans form adjacent to narrow beach deposits and a narrow carbonate shelf. A vertical succession from such a setting would be similar to the basal redbeds and overlying limestones, if the beach deposits were missing.

Massive limestone member

The massive limestone member abruptly overlies the basal red conglomerate member with no evidence of erosion. The massive limestone member is a light- to medium-grey, medium- to thick-bedded, bioclastic packstone which grades into a thin- to medium-bedded, dark grey, bioclastic wackestone upsection (Fig. 18). The limestone typically weathers to a massive, pitted texture, obscuring bedding and any internal features. Bedding is planar to wavy in the massive limestone member and ranges from several centimetres to 3 m in thickness. The limestone is bioclastic packstone in the thick beds (Fig. 19) and a bioclastic wackestone to packstone in the thinner beds (Fig. 20). In places, cleaved surfaces reveal the occurrence of abundant shelly debris. The large bivalve - *Neomegalodus canadensis* (Shimer) (10-20 cm) is common (Fig. 19), particularly in the lower half of the unit, as well as pieces of other bivalves, corals, bryozoa, and possible sponges and algae. The fossils are disarticulated and broken, and in many places are oriented approximately parallel to bedding. Black chert nodules are present in a few beds (Fig. 21). In the central part of the map area the massive limestone is interbedded with well-indurated, quartz-rich sandstone in 20-50 cm thick beds. The megalodontids of the massive limestone member probably lived in shallow water, usually associated with corals and other shallow marine fauna (Hallam, 1981). Thus the massive limestone member is interpreted to be a shallow marine limestone deposit of reworked shelly fauna derived from an inner shelf area with possible small carbonate buildups.

Monotis *limestone member*

The massive limestone member grades up into *Monotis*-bearing limestone (Fig. 18). The *Monotis* limestone member is a 0-14 m thick sequence of yellow and grey, thin bedded micrites with locally common bivalve *Monotis subcircularis* Gabb, and a few thin, calcareous sandstone beds with uncommon symmetrical ripples (Tozer, 1967). The ecology of *Monotis* is not well known. It is generally found in monotypic assemblages in offshore deposits, but appears in a wide variety of marine environments (Silberling, 1985; Ando, 1987). Most workers have suggested that, due to a thin, flat

Figure 19. *Photograph of the massive limestone member of the Tyaughton Formation. Bioclastic packstone with bivalve, coral, and sponge(?) clasts. The large bivalves to the left of the pencil are fragments of* Neomegalodus canadensis *(Shimer). Photograph by P.J. Umhoefer. GSC 1997-53I*

Figure 20. *Photograph of the massive limestone member of the Tyaughton Formation. Wavy bedded bioclastic wackestone near the top of the limestone unit. Tops are to the left. Photograph by P.J. Umhoefer. GSC 1997-53J*

Figure 21. *Photograph of the massive limestone member of the Tyaughton Formation. Wackestone with black chert nodules. Photograph by P.J. Umhoefer. GSC 1997-53K*

shell, it lived attached to floating objects (e.g. Silberling, 1985), but others favour a epifaunal benthic habitat (e.g. Hallam, 1981).

The transition from the massive limestone member to the micritic, *Monotis* limestone member represents a change to a quieter, less energetic environment, perhaps a bay or more offshore position on the shelf. The overall thinning and fining upward in the two limestone members, and the inferred paleoecology of the dominant bivalves, argues that the limestone unit is a transgressive sequence.

Limestone cobble conglomerate member and lower green clastics member

The limestone cobble conglomerate and lower green clastics members are dominantly a ~100 m thick fluvial sequence of conglomerate and crossbedded sandstone (Fig. 22) that abruptly overlies either the *Monotis* limestone or massive limestone member and is best exposed on the ridge 2.5 km northwest of Castle Peak (#2 on Fig. 4).

Where thickest, the limestone cobble conglomerate member varies and includes beds 1-2 m thick of closely packed, subangular to subrounded, grey limestone clasts up to 20 cm across, with a minor matrix component of angular quartz granules (Fig. 23). Locally the clasts are crudely imbricated and the beds are everywhere graded. The lithology of the clasts matches the underlying massive limestone member and a few clasts can be definitively tied to the lower part of that member. Figure 23 shows that some of the largest clasts

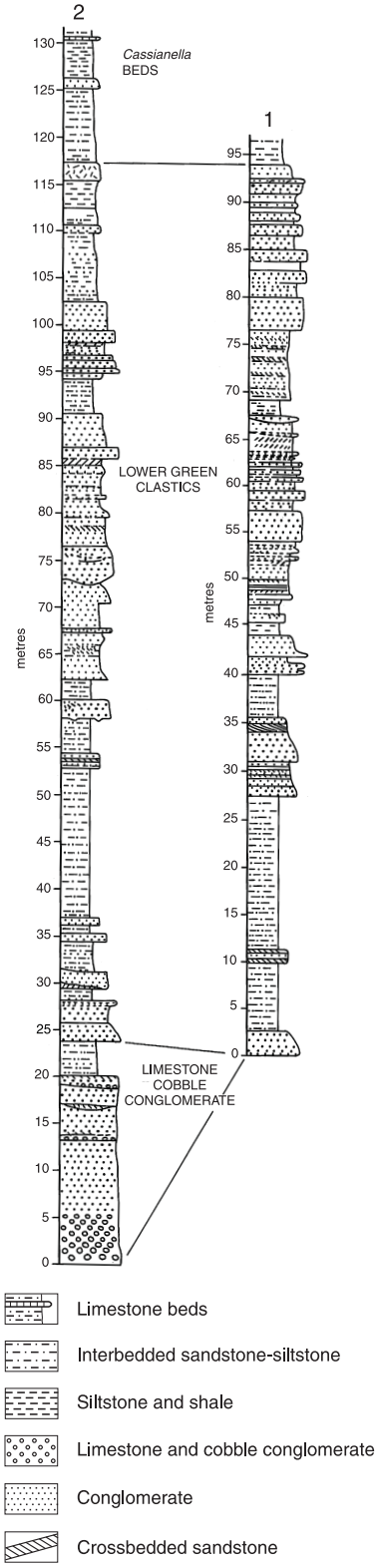


Figure 22. *Measured sections of the limestone cobble conglomerate, lower green clastics, and basal part of the -* Cassianella *beds members. Circled numbers refer to section locations in Figure 4.*

contain neomegalodontid bivalves and one clast has a black chert nodule, both of which are features of the massive limestone member in the same section (cf. Fig. 19, 21).

The basal limestone conglomerate bed grades up into pebble conglomerate (1-8 cm diameter clasts), and then into limestone-pebble sandstone and the sandstone of the lower green clastics member. The sandstone is red- to orange-weathering, comprising angular quartz granules to coarse grained sand. In places where the limestone conglomerate member is missing, limestone-pebble sandstone constitutes the basal few metres of the lower green clastics member (Fig. 22, 24).

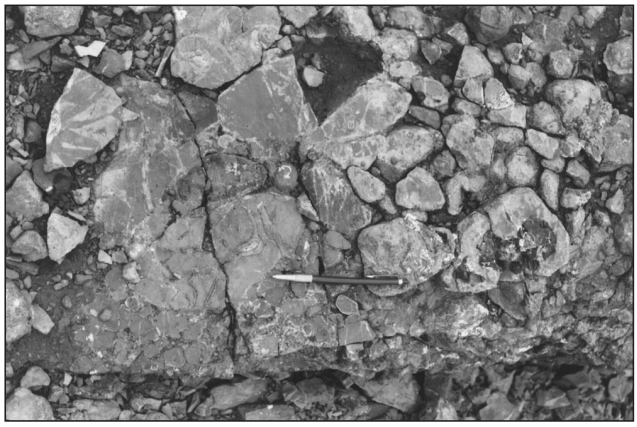


Figure 23. *Photograph of the base of the lowest bed of closely packed limestone-cobble conglomerate member, Tyaughton Formation. Note fragments of the bivalve* Neomegalodus canadensis *(Shimer) in the clasts at the left tip of the pencil and black chert in the clast on the right edge of the photograph, both derived from the massive limestone member below the limestone cobble conglomerate member (Fig. 19 and 21). Photograph by P.J. Umhoefer. GSC 1997-53L*

Figure 24. *Upper part of limestone cobble conglomerate member below lower green clastics member, Tyaughton Formation, where it is gradational up into the lower green clastics member. Photograph by P.J. Umhoefer. GSC 1997-53M*

The main lithology in the lower green clastics member is green, medium- to coarse-grained, lithic sandstone, which is subequally planar laminated and crossbedded (Fig. 25). Tabular and trough crossbedding are both common (Fig. 26). Paleocurrent measurements from cross-strata and a few scour marks (Fig. 27) indicate east to west (present-day co-ordinates) transport of the sand. Green, sandy, pebble conglomerate to pebbly, coarse sandstone make up about 20-60% of the unit. The coarser grained rocks form tabular to

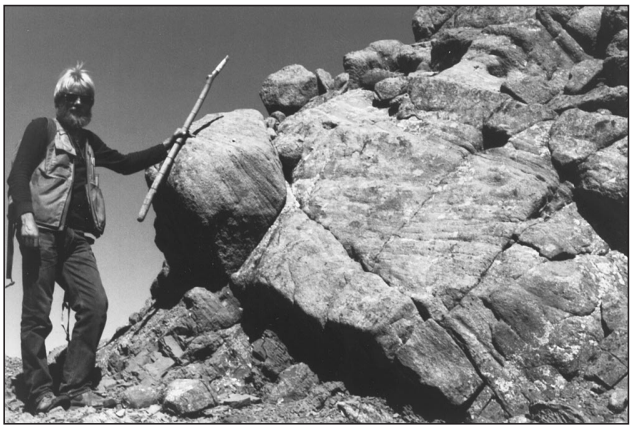


Figure 25. *Photograph of the lower green clastics member, Tyaughton Formation. Thick, channel-fill, pebbly sandstone with large-scale crossbedding. This bed has many linear sole marks as seen in Figure 27. Photograph by P.J. Umhoefer. GSC 1997-53N*

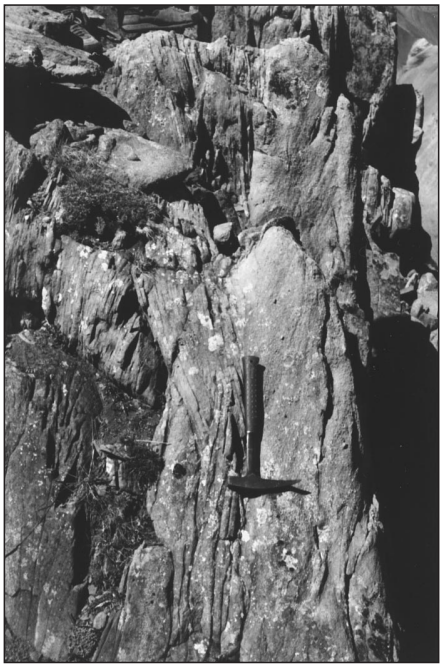


Figure 26. *Photograph of crossbedded sandstone in the lower green clastics member, Tyaughton Formation. Tops are to the right. Photograph by P.J. Umhoefer. GSC 1997-53O*

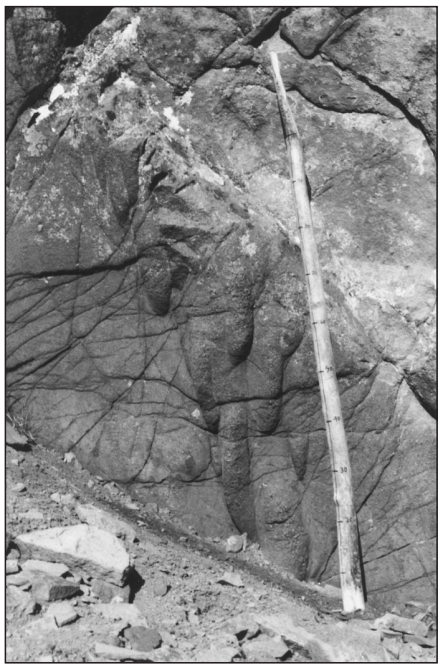


Figure 27. *Photograph of sole marks on the bottom of a bed of the lower green clastics member of the Tyaughton Formation. Photograph is base of bed shown in Figure 25. Scale on stick is in 10 cm intervals. Photograph by P.J. Umhoefer. GSC 1997-53P*

Figure 28. *Gravel bed with preserved dune forms in the lower green clastics member of the Tyaughton Formation. Tops are to the left.Photograph by P.J. Umhoefer. GSC 1997-53Q*

lenticular beds 0.5-2 m thick; a few beds consist of large scale, tabular cross-sets above a scoured base (Fig. 25). Other conglomerate beds have flat, nonerosive bases and comprise poorly sorted pebbles. Conglomerate clasts are subrounded, a few millimetres to 5 cm in diameter, and consist of felsic to intermediate volcanic rocks, limestone, and minor granitoid rocks. Fossils are present only in the uppermost part of the member, transitional to the *Cassianella* beds.

The limestone conglomerate and lower green clastics members are interpreted to be mixed sand and gravel, braided-stream deposits with a transition to marginal marine deposits at the top. The bedding thickness, sedimentary structures, vertical succession of lithologies and structures, and lateral variability are all similar to the Donjek-type braided stream model of Miall (1977). The poorly-sorted conglomerate beds with flat, nonerosive bases, and sinusoidal tops (Fig. 28) may have originated as small debris flows later reworked by river currents. The stratigraphically highest conglomerate and trough crossbedded sandstone may be beach and shoreface deposits. The beds pass upward into 10-20 m of fine- to medium-grained sandstone containing minor, bivalve-bearing beds, and then into the distinctive shell-rich beds of the shallow marine, *Cassianella* beds member. The transition from the lower green clastics to the *Cassianella* beds is similar to a high wave-energy section off the Ventura-Oxnard California coast (McCubbin, 1982), except that the medial Tyaughton Formation is a transgressive sequence whereas the modern California example is a regressive sequence.

Cassianella *beds member*

The dominant lithology in the *Cassianella* beds is green-brown- to brown-weathering, poorly stratified to unstratified, fine grained, calcareous sandstone to sandy siltstone. Trace fossils locally are well preserved in the poorly stratified sandstone, such as the *Rhizocorallium* burrows illustrated in Figure 29. Bivalve-rich, sandy calc-arenite to calcareous

Figure 29. Cassianella *beds member of the Tyaughton Formation. View into a bed of sandstone with* Rhizocorallium *trace fossils. The bed is broken such that the burrows are at low to moderate angles to bedding. Photograph by P.J. Umhoefer. GSC 1997-53R*

sandstone in beds 10-40 cm thick form resistant ribs in the fine-grained sandstone in most places (Fig. 9, 30). The coarse, resistant beds are laterally continuous, at least for hundreds of metres. The calc-arenite beds have a sharp base and are graded in some places. A typical bed has a 10-20 cm thick, basal, shell-rich layer of disarticulated shells, wood, and other organic matter, all oriented parallel to bedding. This layer makes up the entire bed in some cases; in other places, the shell-rich layer grades laterally into calcareous sandstones. The spacing of the shell-rich ribs varies between 2 and 20 m; many sections have a regular spacing of 8-10 m (Fig. 9, sections 1 and 2). Fossils in the shell-rich beds include, most commonly, the robust bivalve *Cassianella lingulata* (Gabb), as well as many species of bivalves and ammonoids (McLearn, 1942; Tozer, 1967, 1979). Above the shell-rich layer, the resistant ribs are typically either massive or parallel-laminated and/or small-scale, ripple-laminated (a few centimetres), medium grained sandstone. The top of this sandstone either grades up into the overlying fine grained sandstone or is sharply overlain by it. Immediately around Castle Peak the *Cassianella* beds member has few to no shell-rich beds. The unit there is a monotonous black to dark brown siltstone to sandy siltstone (Fig. 11).

The physical sedimentology and paleoecology of the - *Cassianella* beds suggest that the shell-rich ribs are storm deposits of shells derived from the slope face interbedded with fair-weather, burrowed, nearshore deposits. The shell-rich beds have most of the features of the storm-deposit model of Kreisa and Bambach (1982) and are similar to calcareous strata of the Middle Triassic Muschelkalk limestones of Germany, interpreted as storm-produced "tempestites" by Aigner (1982). In the storm model (Kreisa and Bambach, 1982), the sharply defined, erosional base and shell layer are deposited during the storm peak; the shells are reworked, but not transported great distances. The upper layer of laminated or massive sands drops out of suspension in the waning stage of the storm; uncommon ripple-laminations signal bed load transport in the final state. The poor stratification of the intervening finer grained beds suggests that they were thoroughly burrowed. The few well preserved burrows such as - *Rhizocorallium* (Fig. 29) are part of the nearshore *Cruziana* ichnofacies which is common in sequences with storm generated beds (Ekdale et al., 1984).

A study of the paleoecology of the Upper Norian Gabbs Formation of western Nevada (Laws, 1982) supports an inner-shelf interpretation for the *Cassianella* beds of the Tyaughton Formation. The middle member of the Gabbs Formation is a bioclastic calc-arenite to calcareous siltstone unit, similar to the *Cassianella* beds. Laws (1982) interpreted five faunal associations in the middle Gabbs Formation, one of which contains some of the same bivalves as in the - *Cassianella* beds of the Tyaughton Formation. This faunal association consists of infaunal (e.g. *Myophoria*) and epifaunal (e.g. *Cassianella lingulata* Gabb) suspension-feeders that live on the inner shelf, probably on a moderately stable substrate (Laws, 1982). Most of the ammonoids in the Gabbs Formation are common to both the Gabbs Formation and *Cassianella* beds of the Tyaughton Formation. Laws (1982) interpreted these as epibenthic scavengers, part of a more offshore paleoenvironment.

Figure 30. Cassianella *beds member of the Tyaughton Formation. Bivalve-rich layer within a typical brown, calcareous, sandstone unit. Striped scale in vertical position is in centimetres. Photograph by P.J. Umhoefer. GSC 1997-53S*

Upper green clastics member

The upper green clastics member is the top unit of the Tyaughton Formation and consists of green, crossbedded, sandy conglomerate to medium grained sandstone (Fig. 9, 31). The unit is missing in some sections, where the Jurassic Last Creek Formation directly overlies the *Cassianella* beds member; elsewhere the upper green clastics member ranges up to 74 m thick and gradationally overlies the *Cassianella* beds.

Fine-pebble conglomerate beds are 10 to 100 cm thick and are locally crossbedded; some grade up into pebbly sandstone. Sandstone beds are green to brown-green, medium- to coarse-grained, and exhibit low-angle, tabular crossbedding, horizontal lamination (Fig. 32), and less commonly, trough crossbedding. Thin, fine-pebble stringers and bivalve-rich layers are locally common in the sandstone. The bivalves are small (a few millimetres to 2 cm) and notably different than most of the species in the underlying *Cassianella* beds. Conglomerate clasts are mainly felsic to intermediate volcanics and minor limestone.

Many aspects of the upper green clastics member resemble well exposed shoreface sequences in southwest Oregon (Bourgeois and Leithold, 1984). The transition from nearshore *Cassianella* beds to shoreface deposits of the upper green clastics is similar to Quaternary sequences along coasts with high wave energy (McCubbin, 1982).

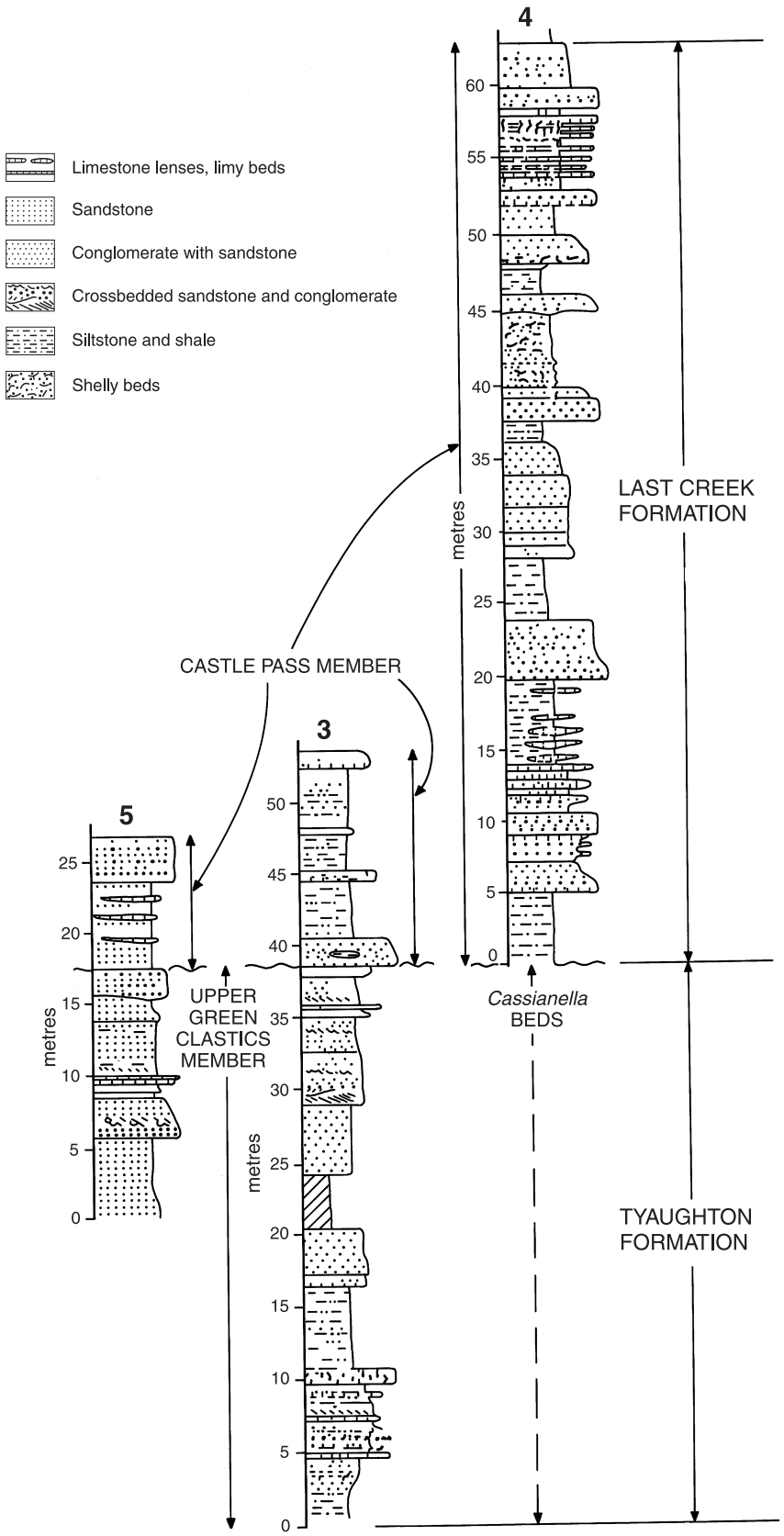


Figure 31. *Measured sections of upper green clastics member, Tyaughton Formation, and basal beds of the Castle Pass member of the Last Creek Formation. Section locations are shown as circled numbers on Figure 4.*

Figure 32. *Upper green clastics member of the Tyaughton Formation. Medium-grained sandstone with low-angle crossbedding. Tops are to the right. Photograph by P.J. Umhoefer. GSC 1997-53T*

Biostratigraphy of the Tyaughton Formation

F.H. McLearn provided early fossil identifications and age determinations for the Tyaughton Formation for C.E. Cairnes (1943). E.T. Tozer undertook a study of the Tyaughton macrofaunas in 1963 and this served as a basis in whole or in part for several publications that involved Late Norian time (Tozer, 1967, 1979). It is from his publications that most information on the biostratigraphy has been derived. M.J. Orchard collected material for conodont studies and from this material and from samples submitted by the writers some additional faunal information has been obtained. This part of this report has been reviewed by Tozer and by Orchard.

The Tyaughton Formation is divided into seven members as described above (Umhoefer et al., 1988). The faunas of these units are listed and discussed from oldest to youngest.

Basal red conglomerate member

This member has yielded rare bivalves which have not provided an age determination (E.T. Tozer, pers. comm., 1987). Conodonts have been recovered from limestone clasts of the conglomerate and these provide a maximum age constraint. These are determined to be late Early Norian (M.J. Orchard, GSC internal report OF-1992-3). The conodont fauna includes the following forms:

Epigondolella triangularis (Budurov)
Epigondolella quadrata Orchard
Epigondolella sp.

The basal red conglomerate member is probably Middle Norian or early Late Norian, although a latest Early Norian age cannot be discounted.

Massive limestone member

This member succeeds the basal red conglomerate, probably across a disconformity. Macrofossils are relatively common but poorly preserved. Only in places where fortuitous weathering provides a reasonably good specimen, can the fossils be recognized. The common, large bivalve *Neomegalodus canadensis* (Shimer) is the most readily recognized but other small, silicified bivalves are present. Corals, echinoid radioles, and one poorly preserved nautiloid have been noted.

One collection by M.J. Orchard from about 5 m below the megalodont interval of the massive limestone member yielded a single specimen of *Epigondolella* sp. and is interpreted as Middle-Late Norian. This is therefore very near the base of the member and no preference for a Middle or Late Norian age is implied. Ichthyoliths were also found in the collection.

A second collection by M.J. Orchard from 5 m below the *Monotis* collection in the overlying member and therefore at or near the top of the massive limestone member yielded two species of conodonts, *Epigondolella bidentata* Mosher, *Epigondolella englandi* Orchard, and one ramiform element. These were interpreted by Orchard as Late Norian. Ichthyoliths were also associated.

The massive limestone member is therefore Late Norian at the top and Late Norian or possibly Middle Norian at the base.

Monotis *limestone member*

The massive limestone member grades up into the *Monotis* limestone member named for the characteristic bivalve, *Monotis subcircularis* Gabb, of early Late Norian age and typically occurring in the Cordilleranus zone. Because of the gradational relation of this member to the underlying massive limestone member they are both considered to be of early Late Norian age in whole or in part; the lower beds of the massive limestone member could be Middle Norian.

The faunas from the *Monotis* limestone member are as follows:

Ammonoids
Halorites cf. *H. americanus* Hyatt and Smith
ammonite indeterminate *(Metasibirites??)*
Bivalves
Monotis subcircularis Gabb
No microfossils have been recovered from the *Monotis* limestone member.

Limestone cobble conglomerate member and lower green clastics member

An erosional disconformity separates the limestone members from the overlying fluvial clastic members and as a result the *Monotis* limestone member below the unconformity may have been removed in whole or in part.

There are no fossils in the fluvial limestone cobble conglomerate except within clasts eroded from the massive limestone member. The overlying lower green clastics member is mainly fluvial sandstone which grades upward into the *Cassianella* beds member and the only fossils found are rare bivalves near the top of the member and a single crinoid at the transition into the *Cassianella* beds. Because of this transition at the top and the disconformity at the base of these two members, they are considered to be closely allied to the *Cassianella* beds in deposition and age.

Cassianella *beds member*

The distinct *Cassianella* beds outcrop as 10 to 40 cm thick resistant ribs of highly fossiliferous sandy limestone. The ribs are regularly spaced in recessive calcareous siltstone and fine grained sandstone. McLearn (1942) first described the rich bivalve fauna and identified the ammonite fauna (Cairnes, 1943) of this member. Tozer (1979) has designated the *Cassianella* beds as the type locality of the Amoenum zone after the occurrence of *Paracochloceras amoenum* Mojsisovics; it is the middle zone of the Late Norian or middle Late Norian.

The nonmarine limestone conglomerate and lower green clastics, and overlying shallow-marine *Cassianella* beds form a transgressive sequence. The base of the transgressive sequence rests unconformably above the thin-bedded – Monotis limestones, therefore it is likely that the entire transgressive sequence (limestone cobble conglomerate, lower green clastics, and *Cassianella* beds) is middle Late Norian (Fig. 7).

The faunas of the *Cassianella* beds are as follows:

Ammonites
- *Paracochloceras amoenum* Mojsisovics
- *Rhabdoceras suessi* Hauer
- *Placites polydactylus* (Mojsisovics)
- *Arcestes* sp.
- *Rhacophyllites* sp.
- cladisctids

Nautiloids
- *Gonionautilus* sp.

Bivalves
- *Cassianella lingulata* Gabb
- *Pecten* cf. *P. yukonense* Lees
- *Pecten tyaughtonae* McLearn
- *Pecten pontiamnis* McLearn
- *Pecten cadwalladerensis* McLearn
- *Pecten* cf. *P. mihanianus* McLearn
- *Pecten* cf. *P. saccoi* Parona
- *Pecten (Variamussium)* sp.
- *Myophoria cairnesi* McLearn
- *Myophoria suttonensis* Clapp and Shimer
- *Myophoria columbiana* McLearn
- *Myophoria adornata* McLearn
- *Myophoria zeballos* McLearn
- *Lima (Plagiostoma)* sp.
- *Lima* sp.
- *Modiola* cf. *M. strigillata* Goldfuss
- *Macrodon (Catella) tyaughtonae* McLearn
- *Pteria* cf. *P. sturi* Bittner
- *Oxytoma* cf. *O. intermedia* Emmrich
- *Plicatula perimbricata* Gabb
- *Enantiostreon* sp.
- *Pinna* sp.
- *Mytilus shulapsensis* McLearn
- *Pleuromya?* cf. *P. elongata* Schlotheim
- *Astarte* sp.

Other macrofossils
- terebratulids
- gastropods (several species)

Conodonts

Conodont collections were made by M.J. Orchard who reported on them in GSC internal report OF-1992-3. The *Cassianella* beds yielded several collections that were characterized by *Epigondolella bidentata* Mosher, *E. englandi* Orchard, and *E. mosheri* Kozur and Mostler and these apparently range through the member. Icthyoliths, foraminifera, and fragments of other faunas were found associated. No radiolarians were noted.

Upper green clastics member

The *Cassianella* beds grade up into the nearshore sandstones and fine-pebble conglomerates of the upper green clastics member. This has been designated the type locality for the Crickmayi zone, the uppermost ammonite zone of the Triassic, and therefore latest Norian (Tozer, 1979). The index species is the heteromorphic ammonite *Choristoceras crickmayi* Tozer (Tozer, 1979). Thin, bivalve-rich beds are locally present in the upper green clastics member. The faunas listed below under ammonoids and bivalves represent several collections by various workers and represent a listing of all known genera and species without any stratigraphic position indicated within the member.

Faunas of the upper green clastics member are as follows:

Ammonoids
- *Choristoceras crickmayi* Tozer
- *Choristoceras* cf. *C. marshi* Hauer
- *Arcestes* sp.

Bivalves
- *Meleagrinella* sp.
- *Modiola* cf. *M. strigillata* Goldfuss
- *Myophoria suttonensis* Clapp and Shimer
- *Tutcheria* sp.
- *Oxytoma* sp.
- pectinids

Conodonts (internal report by M.J. Orchard, GSC internal report OF-1992-3)
- *Misikella posthernsteini* Kozur and Mock (1 specimen)

One sample, possibly from this member yielded icthyoliths, pentacrinoids, gastropods, pelecypods, and ammonoids (GSC internal report OF-1992-3).

Last Creek Formation (new)

The Lower to Middle Jurassic (Upper Hettangian to Lower Bajocian) clastic rocks exposed in the Chilcotin Ranges are herein formally named for the first time. The Last Creek

Formation is best exposed in the vicinity of Castle Peak in the southern Taseko Lakes map area (Fig. 2, 4). The correlative Nemaia Formation is exposed in the Chilko Lake region, in the northeastern Mount Waddington map area. The Last Creek Formation is folded and faulted into slices so that no single section displays the entire formation, but all parts of the unit are represented along Last Creek, a northern tributary of Tyaughton Creek (Fig. 4). O'Brien (1985) used the name Last Creek formation informally for Lower Jurassic rocks ((?)Hettangian and Sinemurian) in the upper reaches of Last Creek, and Umhoefer (1989) extended the unit to include the rocks around Castle Pass, just north of Castle Peak. The total thickness of the Last Creek Formation is impossible to determine. From part sections the lower member is estimated at least 125 m thick and the upper member is at least as thick and probably much thicker. A total thickness of 250 to 400 m is considered reasonable (Fig. 7).

Rocks of this formation were first recognized and mapped by Cairnes (1943, units 8 and 9) but they were described only sparingly. Lower to Middle Jurassic rocks were included in the Tyaughton Formation by Tipper (1978) because the sandstones on either side of the Triassic-Jurassic boundary were thought to be similar. Glover and Schiarizza (1987) and Glover et al. (1988a) also designated the Tyaughton Formation as both Triassic and Jurassic, but the Lower to Middle Jurassic rocks have been separated from the Tyaughton Formation as an unnamed unit on recent maps of the eastern Tyaughton basin (Glover et al., 1987, 1988b; Umhoefer et al., 1988).

Definition of the Last Creek Formation

The Last Creek Formation is Lower to Middle Jurassic (Late Hettangian to Early Bajocian) clastic rocks outcropping in close association with the Tyaughton Formation (Fig. 4). It overlies the Tyaughton Formation across an erosional disconformity. The unit is unmetamorphosed but highly faulted and folded. Individual sections typically represent only part of a single stage but good fossil control around Castle Pass and in the Last Creek and southern Relay Creek valleys, north of Last Creek (O'Brien, 1985; Umhoefer et al., 1988; Fig. 4) allow construction of a composite section (Fig. 7). The Last Creek Formation is interpreted as a transgressive sequence comprising Hettangian to Sinemurian shallow marine, coarse clastic rocks that grade up into Upper Sinemurian to Bajocian deeper marine shales.

The Last Creek Formation is divided into two members, the lower Castle Pass Member and the upper Little Paradise Member (Fig. 33, 34). The Castle Pass Member is a Late Hettangian to Early Sinemurian coarse grained sequence of dominantly sandstone and conglomerate. It is best exposed in the upper part of Last Creek and just northwest of Castle Pass. Near Castle Pass, the unit disconformably overlies the Tyaughton Formation in a steeply dipping, overturned section which is well exposed and contains abundant ammonites (Fig. 4, 11).

The Little Paradise Member is best exposed along the ridge 4 km northwest of Castle Peak (Fig. 4). The unit is dominantly shale, and commonly calcareous, with concretions of centimetre- to metre-scale. Minor thin- to medium-bedded sandstones punctuate the shale sequence.

The base of the Last Creek Formation overlies the Tyaughton Formation across a disconformity, with evidence of a minor erosional episode but little or no angular discordance. The basal Last Creek Formation is broadly similar to the upper green clastics member of the Tyaughton Formation (Fig. 31). The uppermost Tyaughton Formation can be distinguished from the lower Last Creek Formation because the former has a consistent green colour, fewer and finer grained conglomerates (rather than the prominent ribs of conglomerate common in the Last Creek), and a distinctive difference in ammonoid fauna.

The Last Creek Formation is unconformably overlain by the Relay Mountain Group. On a ridge about 3 km west of Relay Mountain, the upper Last Creek Formation (including fossiliferous Lower Bajocian strata) is overlain by the basal Relay Mountain Group (Callovian); Upper Bajocian and Bathonian rocks are missing. Both units have approximately the same strike across the contact and face to the north but they differ in structural style. The Relay Mountain is a homoclinal, north-dipping sequence with minor folds. In contrast, the Last Creek has many close to tight 1 m to 10 m scale folds. Scree on the ridge does not allow detailed examination of the contact itself but the general structural conformity of the two units, combined with the contrasting structural styles at the outcrop scale, suggests that the unconformity between the Last Creek Formation and Relay Mountain Group is angular in detail, and that it represents an episode of deformation (Rusmore et al., 1988).

LAST CREEK FORMATION	MEMBERS	CHARACTERISTIC LITHOLOGY	THICKNESS	AGE
	Little Paradise Member	calcareous black shale, minor sandstone, rare conglomerate	>300 m	Early Bajocian to mid-Early Sinemurian
	Castle Pass Member	volcanic-pebble conglomerate, brown medium- to coarse-grained sandstones, sandy siltstone, rare tuff bands	~150 m	mid-Early Sinemurian to Late Hettangian

Figure 33. *Members of the Last Creek Formation.*

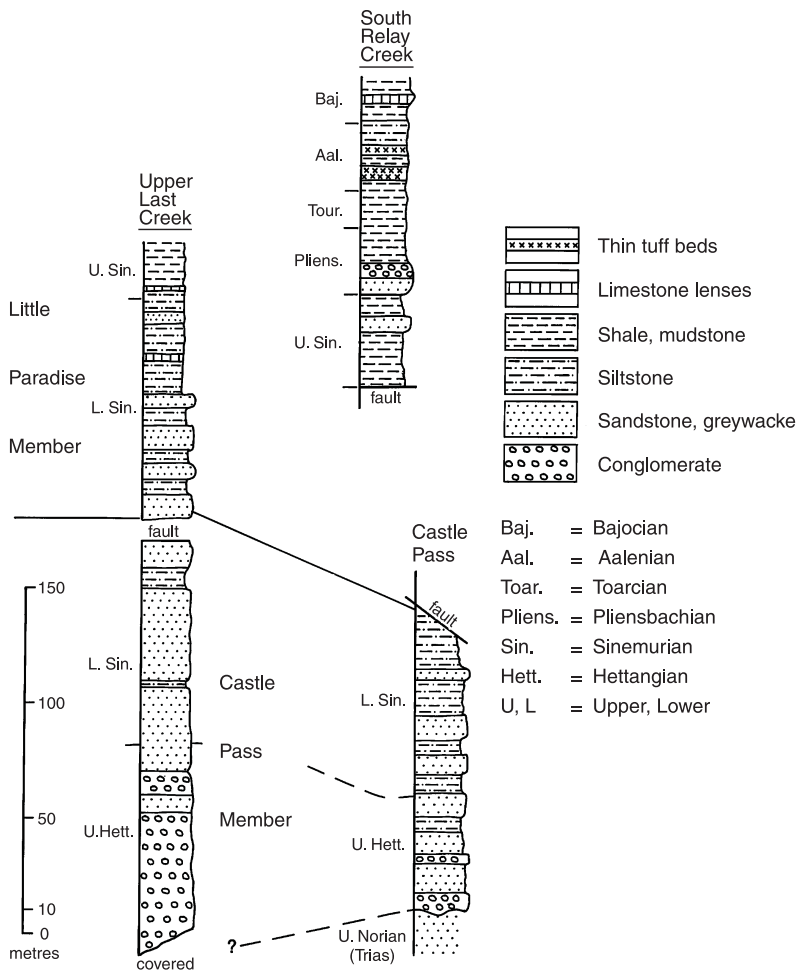


Figure 34.

Schematic sections of Last Creek Formation. Hett. = Hettangian; Sin. = Sinemurian; Pliens. = Pliensbachian; Toar. = Toarcian; Aal. = Aalenian; Baj. = Bajocian.

The conclusion that the upper contact of the Last Creek Formation is an angular unconformity must be viewed with caution, however, because it is only exposed on the one ridge just described and that ridge is part of a large fault-bounded lens along the Castle Pass and Fortress Ridge faults. Schiarriza et al. (1993) map the contact as a high-angle fault. The Castle Pass Fault may have a complex history including oblique-thrust and strike-slip movements entirely after Relay Mountain deposition (Umhoefer, 1989; Schiarizza et al., 1990, 1993). Therefore it is possible that the contrast in the degree of deformation between the Last Creek Formation and Relay Mountain Group, the Castle Pass Fault is due to local complexities in the structural evolution along the Castle Pass Fault and not to an earlier deformation in the Last Creek Formation.

Lithostratigraphy and depositional environment of the Last Creek Formation

Castle Pass Member

The basal member of the Last Creek Formation, the Castle Pass Member (Late Hettangian to mid-Early Sinemurian), comprises interbedded volcanic-pebble conglomerate and brown to green, coarse- to medium-grained sandstone (Fig. 11, 35). The unit becomes finer grained upward with conglomerate dominant in the lower part and sandstone dominant upward. Conglomerate beds are 20-400 cm thick and massive or graded. Conglomerate clasts are dominantly chert, felsic to intermediate volcanic, with minor plutonic and sedimentary clasts; clast diameter ranges up to 8 cm. Carbonized woody debris is common, the largest is a log 15 cm in diameter (O'Brien, 1985). Some conglomerate beds are lenticular channels up to 2 m wide and others change laterally into coarse grained lithic sandstones (Fig. 31).

Sandstone beds are massive, parallel laminated, or low-angle crossbedded, and are commonly graded. Layers of reworked shells are common as thin (5-20 cm), individual beds or at the base of graded beds of conglomerate to pebbly sandstone. The uppermost beds of Hettangian age are distinctive white to pink, plagioclase-rich, fine-pebble to granule conglomerates. Sinemurian rocks consist of brown, calcareous sandstone to sandy siltstone with common calcareous concretions and organic matter. This part of the Last Creek Formation shows a steady change to finer grained rocks upsection. Upper Sinemurian and younger beds of the overlying member are almost entirely siltstone and silty shale. Woody debris and ammonoids are common in the coarse clastic beds of the unit (Frebold, 1967; O'Brien, 1985; Umhoefer et al., 1988). Beds with abundant bivalves, in some beds almost coquinoid in character, are almost entirely restricted to the Castle Pass Member, particularly in lowest Sinemurian beds.

Figure 35. *Type locality of Last Creek Formation. The exposure is sandstone and siltstone of the Castle Pass member, Last Creek Formation. Rocks here exposed are lowest beds of Sinemurian age (upper part of* Canadensis *zone). Photograph by H.W. Tipper. GSC 1997-52B*

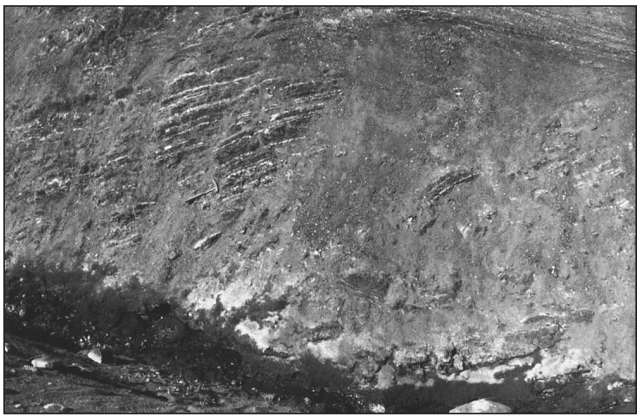


Figure 36. *Little Paradise Member of the Last Creek Formation. Calcareous shales interbedded with clay-rich beds (white) that are possible ash layers. Photograph by P.J. Umhoefer. GSC 1997-53U*

The Castle Pass Member of the Last Creek Formation is a transgressive sequence of nearshore to inner shelf deposits. The conglomerate beds, common woody debris and ammonoid fauna, and low-angle crossbedding found in the basal tens of metres suggest that this sequence was deposited just offshore with a significant fluvial input. The nearshore beds grade up into well bedded sandstone and siltstone probably deposited on an inner shelf below wave base. Uncommon, low- to moderate-angle crossbedding may be part of hummocky cross-stratification, but sedimentary structures are generally uncommon in the upper part of the Castle Pass Member. Original sedimentary structures may have been obliterated by diagenesis or by bioturbation.

Little Paradise Member

The Little Paradise Member of the Last Creek Formation (mid-Early Sinemurian to Early Bajocian) consists of poorly exposed, black, calcareous shales and minor sandstones (Fig. 34, 36). Calcareous concretions are common and range up to 3 m in diameter (Fig. 37). Uncommon, thin, yellow to white, clay-rich layers may be ash beds (Fig. 36). The inner shelf deposits of the Castle Pass Member grade up into siltstone and shale of the Little Paradise Member of the Last Creek Formation which are interpreted to be outer shelf to slope deposits.

The finer grained clastic rocks of this member are in sharp contrast to the coarser grained clastic rocks of the Castle Pass Member. Although the two members are, in general, distinct, there is not a uniformity of lithology within the Little Paradise Member. With adequate fossil control two or three upward-fining sequences may be recognized within the member; without fossils one is only able to assign the strata to this member generally. Some sections show characteristics peculiar to certain time intervals.

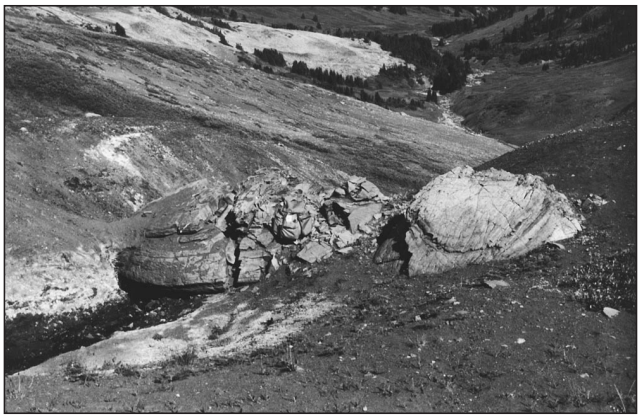


Figure 37. *Metre-scale calcareous concretions in the Little Paradise Member of the Last Creek Formation. Photograph by P.J. Umhoefer. GSC 1997-53V*

Above the Castle Pass Member, the mid-Lower Sinemurian to uppermost Sinemurian strata are mostly black shale to sandy siltstone. The beds are finely and uniformly laminated in 1-3 cm bands. Fine sedimentary structures are, for the most part, absent or rare. Some sandier beds are graded and some beds are calcareous. These sandstones may be turbidites. The Upper Sinemurian beds may be massive shale or silty shale in places with no stratification; limy, brown-weathering concretions to 10 to 15 cm are distributed erratically and commonly contain ammonites, either as one or two specimens or a packed concentration of specimens.

The Pliensbachian rocks have not been found in contact with the Sinemurian beds except at the head of South Relay Creek, northwest of Last Creek. Here upper Lower Pliensbachian conglomerate and pebbly sandstone or grits appear to rest unconformably on Lower Sinemurian beds, probably an erosional disconformity. These pass abruptly into black shale or silty shale of the Upper Pliensbachian and Toarcian but a continuous section was not seen. Elsewhere, north of Castle

Peak, Lower Pliensbachian coquinoid sands were noted and are slightly older than the aforementioned coarse clastic rocks. The Toarcian is typically shaly with a few beds replete with grey, limy concretions, usually without fossils (Fig. 36).

Aalenian shales are commonly finely laminated with laminae up to 1 cm thick; some laminae are probably fine air-fall tuff. Some thin, limy bands are also present. The relation to overlying or underlying beds is unknown but no discordance is indicated.

The Bajocian rocks are a typical shale sequence with rare silty beds, possibly a few fine tuffaceous beds, and one or two beds of white-weathering, grey limestone up to 3 m thick, possibly slightly more. Ammonites are present but generally are not abundant. The shales are not markedly well bedded nor laminated but attitudes are determined from differentially weathered beds.

Variations in the lithology of the Little Paradise Member suggest slight changes or interruptions in deposition. When considered in conjunction with the Castle Peak Member, it is apparent that there are at least two upward fining sequences in the Last Creek Formation, the first deposited from Late Hettangian to Late Sinemurian and one from late Early Pliensbachian to Early Bajocian (Fig. 7). Poor exposure prevents recognition of finer scale sequences but it is entirely possible that minor repetitive interruptions in sedimentation have produced a cyclical style of upward fining sequences.

Biostratigraphy

The Last Creek Formation has a diverse ammonite fauna, of which only a small part has been systematically studied. The following account of the fossil faunas is extracted from many reports and unpublished studies, some of which are ongoing. The information on the Hettangian beds is from current work by Tipper and published work by Frebold (1967). Sinemurian and Pliensbachian faunas were briefly studied by Frebold but only internal reports resulted. Tipper is continuing these studies. G.K. Jakobs has identified the Toarcian faunas as part of her doctoral thesis studies (1992). The Aalenian faunas were studied and published by Poulton and Tipper (1991). The Bajocian faunas are being studied by Poulton and the information included here are a result of his investigations.

Castle Pass Member

The faunas of this member are the most abundant and most diverse of any part of the Last Creek Formation. In general they are well preserved, three dimensional forms and include ammonites, bivalves, gastropods, rare brachiopods, and rare coleoids. The ammonites of the member can be organized into three or four assemblages or zones and these are currently being studied with the view of establishing a North American ammonite zonation for the Sinemurian.

The lowest zone or assemblage is of Late Hettangian age and is restricted to the basal pebble conglomerate beds of the formation. For the most part the taxa are small (3 to 5 cm in diameter), but rare forms reach 25 cm or more. Broken shell fragments indicate that there were many large forms but the high energy environment of the conglomerates have destroyed the large forms; the small forms survived in small lenses, in thin sandy beds, or in the interstices of the coarser conglomerates.

The following ammonite fauna has been recognized:

Sunrisites sunrisensis Guex
Sunrisites spp. (probably new species)
Badouxia n.sp. (possibly more than one)
Schlotheimia sp.
Alsatites? sp.
Eolytoceras sp.
Pseudaetomoceras sp.
Paracaloceras sp.
phylloceratids

Immediately above the conglomerate beds in a sequence that is mainly greywacke and sandy siltstone, an abundant fauna has been obtained by several workers and was studied by Frebold (1967). This is the basis of his Canadensis zone, the only established North American Sinemurian zone. The fauna is abundant, three-dimensional, well preserved, diverse and contains several forms that are endemic to the Pacific region. Whereas Frebold (1967) considered the Canadensis zone to be Late Hettangian, the most recent interpretation (Pálfy, 1991) places the lower part of the zone in the Hettangian and the upper part in the Sinemurian. The base of the zone is marked by the incoming of *Badouxia canadensis* (Frebold) and the beginning of Sinemurian time is generally the incoming of *Badouxia columbiae* (Frebold). The age of the zone is controversial and further work is needed to establish the Hettangian-Sinemurian boundary.

The Hettangian part of the Canadensis zone does not have a diverse fauna and the ammonite fauna present is as follows:

Badouxia canadensis (Frebold)
Badouxia occidentalis (Frebold)
Badouxia n.sp.
Metophioceras rursicostatum (Frebold)
Angulaticeras marmoreus (Oppel)
Eolytoceras sp.
Paracaloceras sp.

One or two beds have an abundance of *Badouxia canadensis* Frebold to the exclusion of all other genera or species. Bivalves are common but generally small.

The earliest Sinemurian fauna is the upper part of the Canadensis zone fauna. It is diverse and abundant and generally well preserved. Of the following forms most were described by Frebold (1967) and a few were collected later by Tipper and are specifically undescribed.

Ammonites:
Badouxia columbiae (Frebold)
Badouxia canadensis (Frebold)
Badouxia occidentale (Frebold)
Badouxia sp. (new)
Metophioceras rursicostatum (Frebold)
Angulaticeras sp.
Eolytoceras tasekoi Frebold
Eolytoceras cf. *E. tasekoi* Frebold

Eolytoceras sp. indet.
Phylloceras s.l. sp.,
Canavarites(?) sp.
Alsatites(?) sp.
ammonites gen. et sp. indet.
Bivalves
Weyla sp.
trigoniids
others as yet not described
Gastropods and brachiopods noted but not common

The Canadensis zone fauna has been a stratigraphic problem from the time it was studied by Frebold (1967). Endemic genera and species comprise the more common forms and genera or species that would indicate a definitive stratigraphic position for the fauna in terms of European zonation are either poorly preserved, not present, few in number of specimens, or not found in place. When the fauna is compared to the Nevada fauna (Guex, 1980) or Queen Charlotte Island fauna (Pálfy, 1991; Tipper and Guex, 1994) it is clear that the fauna of the underlying beds is late Hettangian. Relation of the beds in the upper part of the zone with the overlying beds is more questionable; the contact is everywhere a fault but the amount of missing section, if any, is negligible. The upper part of the zone where exposed in Queen Charlotte Islands clearly has *Badouxia columbiae* associated with Sinemurian forms such as *Vermiceras* sp. and *Metophioceras* sp. aff. *M. rotarium* (Pálfy, 1991). The Hettangian-Sinemurian contact will remain controversial until additional data are available. A restudy of the Last Creek faunas and further collecting may give a clearer understanding of the problem.

The uppermost fauna of the Castle Pass Member is present at the head of Last Creek and at Castle Pass (Fig. 4). The sandy beds have a diverse fauna, well preserved and abundant characterized by arietitids such as *Coroniceras*. The forms in general are large; some specimens suggest diameters of 40 cm or more. Practically all are three-dimensional with little or no secondary compression. The fauna has not been studied in detail but preliminary examination suggests that the fauna correlates mainly with the Bucklandi zone and the lower half of the Semicostatum zone of northwest Europe, the earliest Sinemurian zones. The fauna recognized is as follows:

Ammonites
Coroniceras sp.
Arietites bisulcatum
Tmaegoceras sp.
Arnioceras sp.
Primarietites(?) sp.
Agassiceras(?) sp.
Vermiceras(?) sp.
Boucaulticeras(?) sp.
Megarietites(?) sp.
Bivalves, gastropods, and brachiopods are common.

Some of these forms are recognized in Queen Charlotte Islands (Pálfy, 1991) and in Nevada (Taylor, in press) and the fauna occupies the same stratigraphic position, above the Canadensis zone and below the next obvious faunal assemblage characterized by species of *Arnioceras*.

Little Paradise Member

The faunas of this member begin stratigraphically above the arietitid faunas of the highest beds of the Castle Pass Member; the contact is poorly exposed. The lowest fauna is characterized by species of *Arnioceras* in relative abundance and are well-preserved and generally small - less than 7 cm in diameter. The fauna can be found at the head of Last Creek, at the head of Relay Creek, and at Castle Pass (Fig. 4). The fauna, although abundant, is not diverse. Ammonites recognized are *Arnioceras miserabile*, *Arnioceras arnouldi(?)*, *Arnioceras* spp., arietitids not determined as yet, and phylloceratids. Small bivalves are associated in a few beds.

An unusual fauna was found on the north side of Castle Pass in a limestone lens. It is remarkable in that it has not been reported widely in North America but in this locality ammonites are abundant, well preserved, diverse, and generally large – up to 30 cm in diameter. The presence of *Caenisites turneri* would argue for correlation with the Turneri zone of northwest Europe, the youngest zone of the Early Sinemurian and would suggest a position for the fauna at Castle Pass above the *Arnioceras* assemblage described above. At the head of Last Creek, a small collection was made a few metres above the *Arnioceras* fauna and a few ammonites in this collection are similar to some ammonites in the collection from Castle Pass. Tentatively, because of the faunal similarity and the stratigraphic position, these two collections are considered to be the youngest Early Sinemurian in the area.

This late Early Sinemurian fauna has not been studied in detail but a preliminary examination suggests the following ammonites are present:

Caenisites turneri (J. de C. Sowerby)
Coroniceras(?) spp.
Aegasteroceras(?) sp.
Arnioceras sp. aff. *A. arnouldi* (Dumortier)
Hypasteroceras(?) sp.
Partschiceras(?) sp.
lytoceratid
several forms generically not determined, possibly new
Associated with the collection in Last Creek are a few coarse bivalves.

Faunas characteristic of Late Sinemurian time are rare but there is sufficient data to suggest the rocks of this age are mainly shale or siltstone and shale. Mainly ammonites have been found. A single specimen of *Asteroceras* sp. and an oxynoticeratid, possibly *Gleviceras* sp., were found *ex situ*. Near the head of Relay Creek and in a small exposure south of Spruce Lake *Paltechioceras* spp. are common in small limy nodules in shale. In some beds south of Spruce Lake the bivalve *Posidonotis semiplicata* (Hyatt) was found and may mark the top of the Sinemurian (Pálfy, 1991). Although faunas are sparse, what evidence there is, suggests the presence of a nearly complete Upper Sinemurian section.

Pliensbachian time is poorly represented. The oldest ammonite fauna comprises *Acanthopleuroceras* sp., *Metaderoceras* sp., *Tropidoceras* sp., *Reynesocoeloceras* sp., *Gemmellaroceras(?)* sp., and a phylloceratid.

Stratigraphic relations to other faunas is unknown in this area but this fauna most probably represents the Whiteavesi zone, mid-Early Pliensbachian.

A slightly younger Early Pliensbachian fauna representing the latest part of the Freboldi zone, ie. latest Early Pliensbachian, was found in a conglomeratic sequence that may rest directly on Lower Sinemurian beds. The fauna comprises *Dubariceras freboldi* Dommergues, Mouterde and Rivas, *Metaderoceras* sp., *Oistoceras* sp., *Reynesocoeloceras* sp., *Fuciniceras* sp., *Fanninoceras* sp., *Phricodoceras* sp., *Reynesoceras* sp., and several phylloceratids. This is a fauna transitional from Early to Late Pliensbachian.

One or two small exposures of shales have yielded a few fragmentary Late Pliensbachian fossils. These include *Fanninoceras* sp. and *Protogrammoceras* sp.

The Toarcian shales are not exposed in section but several isolated localities with ammonites suggest that much, if not all, of Toarcian time is represented. The ammonites were studied by G.K. Jakobs who reported the presence of *Cleviceras* cf. *C. chryanthemum* (Yokoyama), *Dactylioceras* cf. *D. commune* (Sowerby), *Phymatoceras* cf. *P. rude* (Simpson), *Lytoceras siemensi* (Denckmann), *Phymatoceras* cf. *P. hillebrandti* Jakobs, *Dumortieria? phantasma* Jakobs and *Hammatoceras* sp. (Jakobs, 1992). Preservation is, in general, poor and specimens are few.

The Aalenian beds are mainly or entirely a shale-siltstone sequence. Some beds are finely laminated and the lighter coloured laminae could be fine tuffs. Thin limy shale beds up to 2 cm thick occur sparingly. The faunas are locally abundant and generally compressed and include *Tmetoceras scissum*, *T. kirki*, *Erycitoides* sp. aff. *E. howelli*, *E. kialagvikense*, *Planamatoceras* sp., *Zurcheria* sp., and *Holcophylloceras* sp. Such forms indicate Early and Middle Aalenian but as the section is tectonically complex and there is nothing to suggest a discordance above or below these beds, a complete section may be present. A more complete account has been published by Poulton and Tipper (1991, p. 13-16).

The Bajocian shale-siltstone sequence is the uppermost part of the Last Creek Formation, Little Paradise Creek Member and has a diverse but sparse ammonite fauna. No section is known where a complete biostratigraphic sequence is present. The genera and species listed here represent, for the most part, single isolated specimens and only a few beds have two or more species or genera associated. Bivalves and belemnites are present but extremely rare. The ammonite fauna suggests that all Early Bajocian time may be represented in this shale-siltstone sequence and two collections suggest the possibility of Late Bajocian time being represented in places. However, except for one brief paleontological report (Frebold et al., 1969), no paleontological report has been published. A discussion of the biostratigraphy of the Bajocian is included in a report by Poulton (1997) from which the following lists of faunas is largely derived.

Ammonites
- *Chondroceras marshalli* (McLearn)
- *Chondroceras* sp.
- *Stephanoceras (Skirroceras)* cf. *S. kirschneri* Imlay
- *Stemmatoceras* sp. indet.
- *Stephanoceras* sp. indet.
- *Witchellia(?)* sp.
- *Holcophylloceras* cf. *H. costisparsum* Imlay
- *Oedania?* sp. or *Pseudolioceras?* sp.
- *Sonninia* sp.
- *Kumastephanus* sp.
- *Asthenoceras* sp.
- *Phylloceras?* sp.
- *Lissoceras?* sp.
- *Fontannesia?* sp.
- *Dorsetensia?* sp.

Bivalves
- *Myophorella yellowstonensis* Imlay
- *Inoceramus* sp.
- ostreiids
- pectinaceans

Other macrofaunas of the Last Creek Formation

Although ammonoids are by far the most abundant fauna of the Last Creek Formation, there are other forms that are significant. These will be briefly mentioned relative to the ammonoids.

Bivalves. In the coarse Hettangian beds at the base of the formation, generally small bivalves are found, commonly broken, generally thin-shelled, but fairly diverse. Pectinid forms are common. Most are abraded and suggest they have been transported. They have not been studied. Species of Late Hettangian *Weyla* are present, the oldest known occurrence of this genus (Damboranea, 1987).

In the basal Sinemurian beds, one or two specimens of *Weyla* have been noted. These are in the upper half of the Canadensis zone and are associated with *Badouxia columbiae* (Frebold).

In Lower Sinemurian beds bivalves, trigoniids, pectenids, and others are associated with *Coroniceras* and *Caenisites*. These are commonly thick-shelled and moderately large.

In Lower Pliensbachian beds several large *Weyla* specimens are found as well as a few other bivalves. In the Toarcian, Aalenian, and Lower Bajocian beds, bivalves are rare and comprise single thin-shelled pectenids and other small, smooth forms.

Gastropods. In the Upper Hettangian rocks and in lower beds of the Lower Sinemurian are the only places gastropods have been found. These are generally high-spired forms up to 4 or 5 cm in length.

Coleoids. Coleoids are rare but a few specimens of *Atractites* are scattered through the Lower Sinemurian strata. Only in one Lower Pliensbachian bed is there an abundance of specimens of *Atractites?* but these are poorly preserved. In mid- to Upper Toarcian beds and Lower Bajocian beds a few belemnites were noted and were reported on by J.A. Jeletzky (GSC internal report Km-10-1978-JAJ).

1) *Belemnopsis* sp. indet.
Nannobelus-like representative of Belemnitidae
These are associated with the ammonites *Phymatoceras* sp. and *Dactylioceras* sp. of the Toarcian.

2) *Dicoelites* (s.l.) n. sp.
indeterminate true belemnites
Probably are mid- to Late Toarcian

3) *Dicoelites* (s.l.) n. spp.
?Belemnites s.s. (=*Passaleuthis*)
Occurs with *Polyplectus* sp. (Toarcian)

4) *Acrocoelites* sp. indet. cf. *Dactyloteuthis* sp. indet.
Associated with early Bajocian *Chondroceras?* sp. and a sonninid

Microfossils. Little work has been attempted on microfossils of the Last Creek Formation. Several samples were digested in an attempt to recover radiolarians but so far nothing has been recovered. Foraminifers or ostracodes have not been found. In two samples collected by M.J. Orchard from beds of Late Hettangian and/or earliest Sinemurian a few ichthyoliths were recovered.

Nemaia Formation (new)

The Lower and lower Middle Jurassic rocks exposed in the northeastern part of the Mount Waddington map area form a distinct sequence of volcanogenic clastic rocks including sandstone, siltstone, shale, minor tuff, sandy limestone, and fine-pebble conglomerate. The newly designated Nemaia Formation is a conformable sequence, as far as has been determined, including Lower Sinemurian to (?)Upper Bajocian rocks, but fossils are only locally common and paleontological control is barely adequate (Fig. 38). The Nemaia Formation differs from the Last Creek Formation in the dominantly coarse grained nature of the upper part (Bajocian) of the Nemaia Formation and fine grained lowest beds (Lower

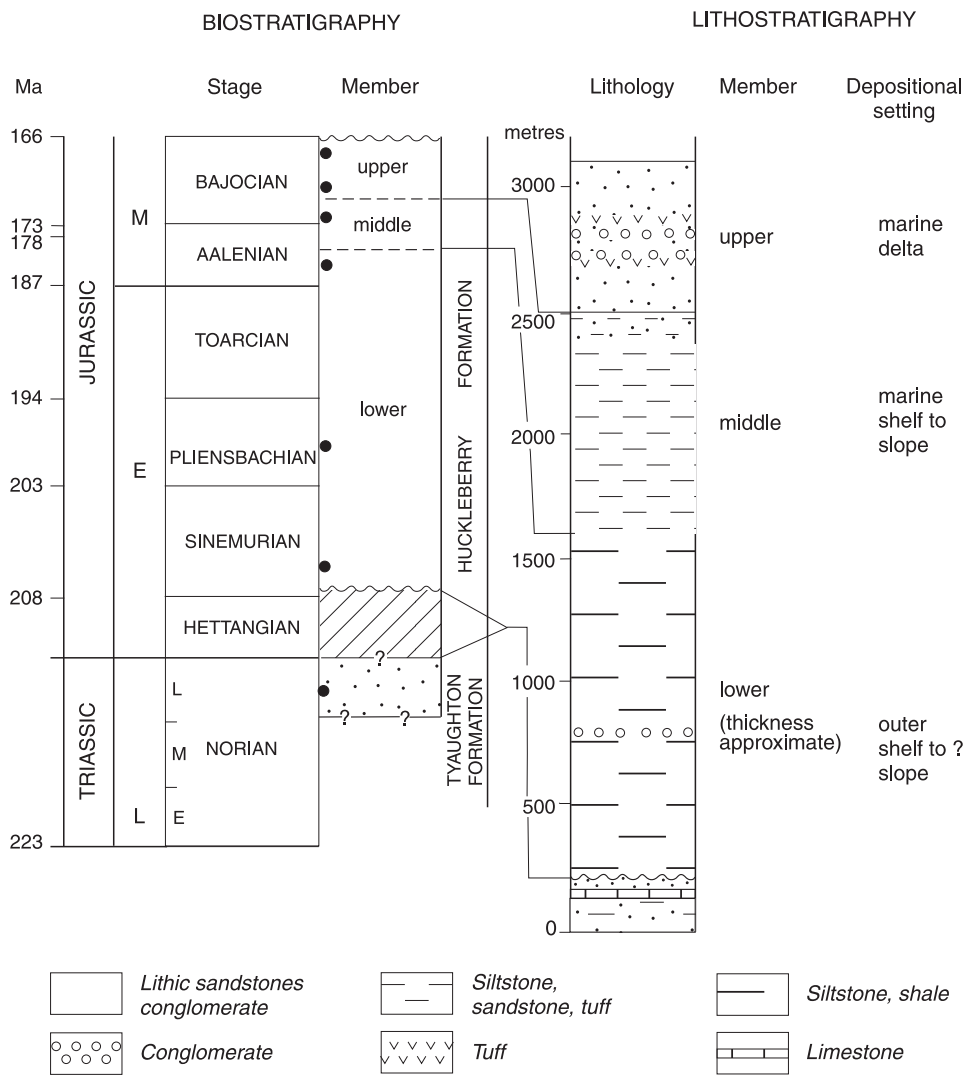


Figure 38. *Biostratigraphic and lithostratigraphic columns of Nemaia Formation, Chilko Lake area.*

Sinemurian); both units are fine grained and broadly similar in strata of Early Sinemurian, Pliensbachian, Toarcian, and Aalenian age (compare Fig. 7, 38).

The Nemaia Formation is found in many fault-bounded blocks between the Yalakom and Tchaikazan faults from Konni Lake to Tatlayoko Lake (Fig. 2, 5), but most of these are below tree-line and poorly exposed. The unit is well exposed on Huckleberry Mountain and all along the Nemaia Range and most of our observations are from those areas (Fig. 5, 39, 40). The unit includes 100 m to kilometre scale, open to closed folds, and a few minor, high-angle faults, but overall both of the main outcrop areas are conformable sequences that face north. Because the two areas lie across Chilko Lake from one another, it is possible they are simple lateral equivalents or offset a few kilometres dextrally along the fault that truncates the north side of the Huckleberry Mountain outcrop area and the southwest side of the Mount Nemaia outcrop area (Fig. 5).

Definition of the Nemaia Formation

The Nemaia Formation is considered to be one conformable unit at least 3 km thick (Fig. 38), but in no single location is the whole unit exposed with adequate fossil control to confirm this conclusion. The upper part of the formation (Aalenian to (?)Late Bajocian) is conformable in many sections.

The entire unit is divided into three members (Fig. 38, 41). The type section on Huckleberry Mountain (Fig. 39, 42, 43) exposes the upper part of the Lower Member, and the Middle and Upper members in a section about 1.5 km thick (Fig. 38; 44, in pocket; 45). The Lower Member is dominantly shale, siltstone, and minor sandstone and conglomerate. The uppermost beds of the unit are of Aalenian age and comprise black to dark grey, finely laminated and banded shales with a few thin interbeds of lithic sandstone and rare brown-weathering calcareous lenses. Sedimentary structures are rare. The best exposures of the Aalenian strata are high on the south side of Huckleberry Mountain (Fig. 39). Unfossiliferous beds on the southern flank of Mount Nemaia are probably correlative.

In Lincoln Creek, 4 km north of Tatlayoko Lake (Fig. 5), a short section of Early Pliensbachian age consists of siltstone, sandstone, and grit to fine-pebble conglomerate. The section varies from outcrops dominated by shale and siltstone to outcrops dominated by coarse grained sandstone. Sandstones are typically medium- to thick-bedded, arkosic to lithic, medium grey, and do not display many sedimentary structures. Rocks broadly similar to those along Lincoln Creek, but lacking fossils, are present on the south side of the Mount Nemaia ridge, low in the section. Recently Paul Schiarizza of the British Columbia Geological Survey collected an ammonite *Phricodoceras* sp. of Early Pliensbachian age from an exposure on the east side of Chilko Lake near the mouth of the creek from Konni lake. This genus is rare in North America (Fig. 5).

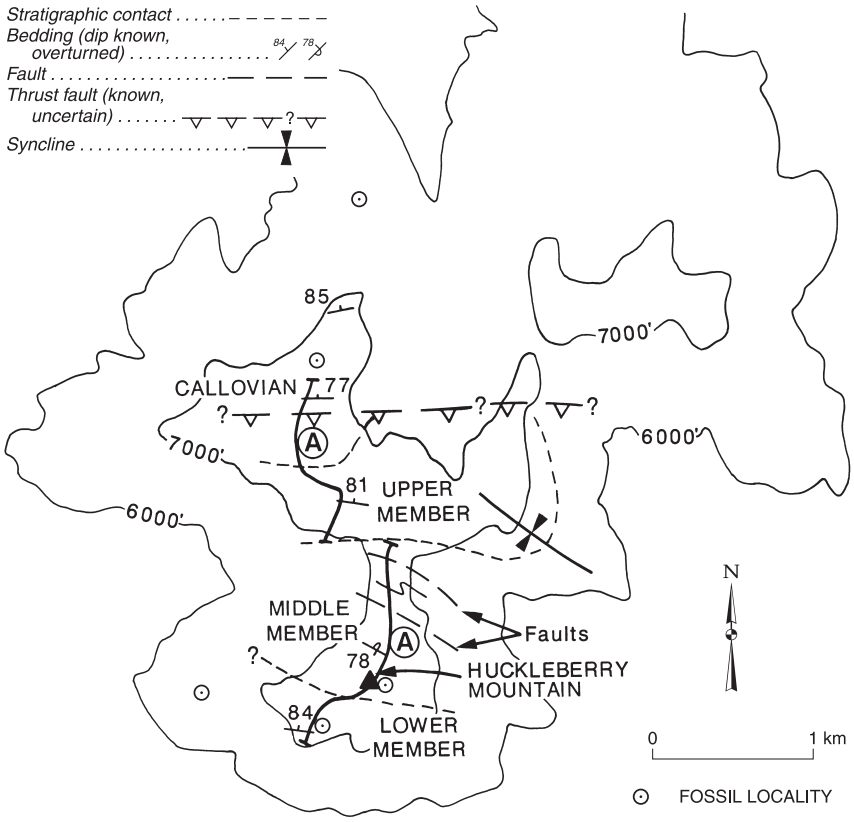


Figure 39. *Simplified geological map of Huckleberry Mountain (see Fig. 5 for location). Circled A is the route of the measured section (Fig. 44).*

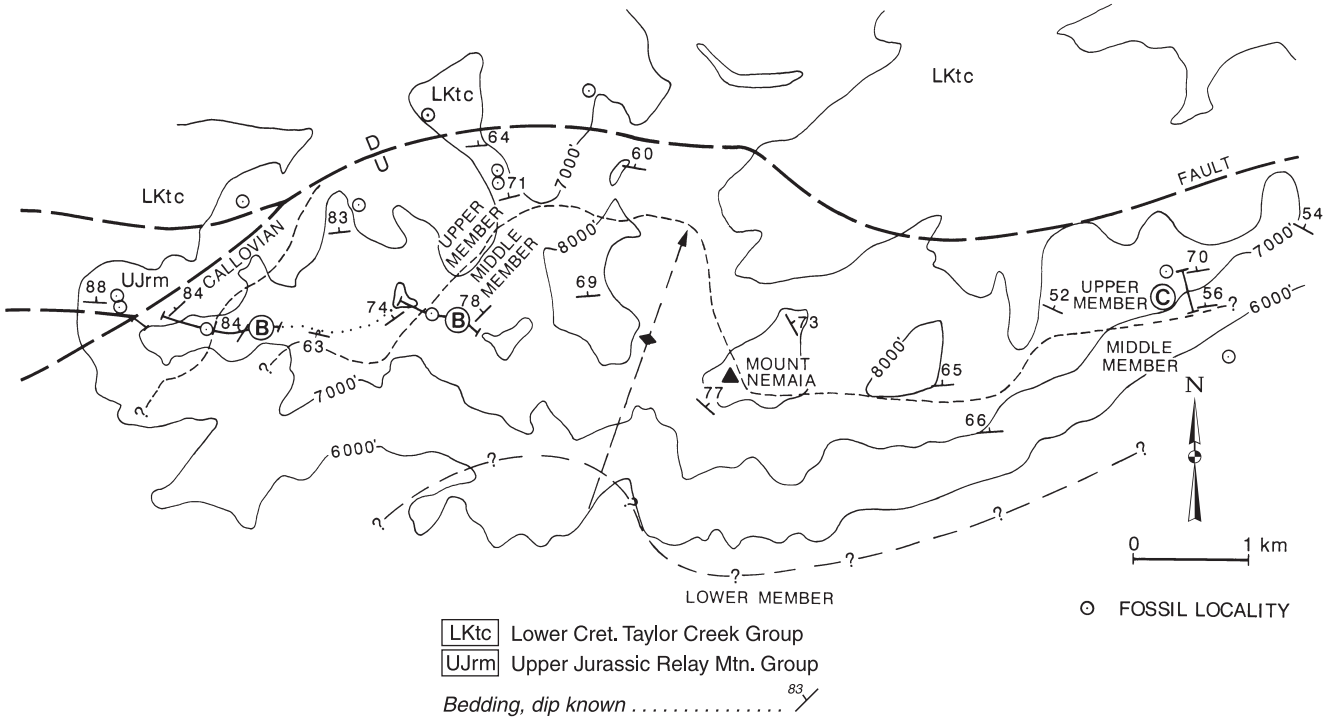


Figure 40. *Simplified geological map of Nemaia Range (see Fig. 5 for location). Circled B and C are routes of measured sections (Fig. 45 and 47, respectively).*

	MEMBER	CHARACTER LITHOLOGY	THICKNESS	AGE
NEMAICA MOUNTAIN FORMATION	UPPER	lithic, sandstone and conglomerate, siltstone, minor tuff	450-650 m	Early Bajocian
	MIDDLE	siltstone and fine grained sandstone, tuff(?)	~900 m	earliest Bajocian
	LOWER	black siltstone and shale, local volcanogenic sandstone and conglomerate	>1 km?	Early Sinemurian to Aalenian

Figure 41. *Members of Nemaia Formation.*

On the north side of Mount Tatlow and south of Konni Lake (Fig. 5), in the Taseko Lakes area, a fault bounded exposure of Lower Sinemurian rocks consists of black to dark grey shale and siltstone with greenish tuff(?) or volcanogenic sandstone and minor calcareous beds. A few fossils indicate an Early Sinemurian age. Riddell et al. (1993) mapped those rocks as the Last Creek Formation. We include these rocks in the lower Nemaia Formation because similar strata occur across the Nemaia valley, near the base of Mount Nemaia, but are unfossiliferous. The unfossiliferous shales low on Mount Nemaia appear to be stratigraphically continuous with the younger strata high on the ridge.

The base of the formation is nowhere exposed. The Lower Member apparently is structurally conformable with the underlying Tyaughton Formation, as defined herein, just east of the northern end of Tatlayoko Lake (Fig. 5), but poor exposure prohibits knowledge of the contact itself. The belt of Lower Member south of Konni Lake is stratigraphically underlain by Hurley Formation (Riddell et al., 1993).

Conformably overlying the Lower Member is the Middle Member; the transition is from shale to siltstone over a few metres. The middle member is typically siltstone and shale, minor fine-grained sandstone, and less common coarse grained sandstone. It varies from recessive to more siliceous, resistant ribs. The unit is best exposed on the south side of Huckleberry Mountain, on Mount Nemaia (Fig. 46), where it forms the high peaks and on the road at the north end of Tatlayoko Lake. Fossils are scarce, but a few sonninid ammonites suggest a basal Early Bajocian age.

Conformably overlying, and abruptly gradational with the Middle Member is the coarse grained Upper Member comprised of medium- to coarse-grained sandstone and

Figure 42. *Aalenian shales of the Lower Member of Nemaia Formation, on Huckleberry Mountain, southwest side. Photograph by H.W. Tipper. GSC 1997-52C*

Figure 43. *Middle and uppermost part of the lower members of the Nemaia Formation. View looking south at Huckleberry Mountain and the ridge north of it. The lower-middle member contact is just south of the top of Huckleberry Mountain. Photograph by P.J. Umhoefer. GSC 1997-53W*

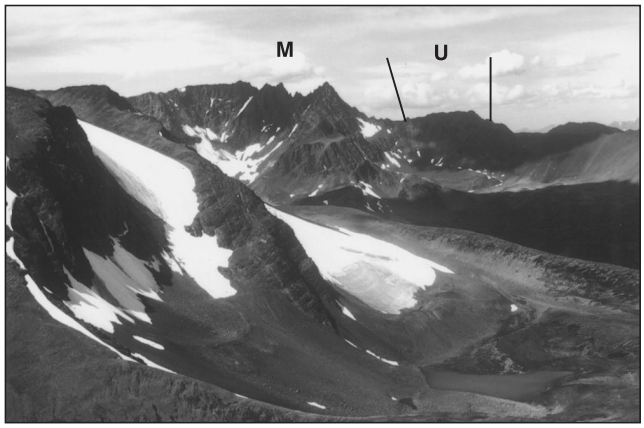


Figure 46. *View to west of north ridge of Mount Nemaia; Middle Member (M) of Nemaia Formation (horizon ridge) overlain by Upper Member (U). Photograph by H.W. Tipper. GSC 1997-52D*

fine-pebble conglomerate interbedded with laminated silty shale and centimetre-scale turbiditic sandstones. Exposure varies from a minimum of 450 m to greater than 600 m thick, and is most likely at least a kilometre thick based on partial sections examined on the eastern and western ridges of Mount Nemaia. The Middle and Upper members form a coarsening-upward sequence.

The Upper Member (and the formation) is overlain by Upper (?)Bathonian-Callovian strata containing sandstone, siltstone, and fine-pebble conglomerate (Tipper, 1969). On Huckleberry Mountain the Upper Member is thrust(?) over a distinctly finer grained section of Callovian shale and siltstone at least 400 m thick (Fig. 39). In contrast, on the western ridge of Mount Nemaia, the Upper Member is overlain abruptly by an additional 340 m fining up sequence of sandstone and volcanic-pebble conglomerate interbedded with subordinate laminated siltstone. This section is generally similar to the Upper Member of the Nemaia Formation, but lacks the common mud ripups and carbonaceous material found in the Upper Member on Huckleberry Mountain. Abruptly and conformably above the latter section are 25 m of siltstone and shale, that contain the Bathonian-Callovian ammonite *Lilloettia*; rare carbonate beds are also present.

Where the (?)[Upper Bathonian]-Callovian rocks are well-exposed above the Nemaia Formation on western Mount Nemaia, the units are conformable. The lack of Upper Bajocian and Bathonian fossils suggests that a significant period of nondeposition is represented by the contact, but poor fossil control renders this a tentative conclusion.

Definition of the members of the Nemaia Formation

Lower member

Fine grained clastic rocks characterize this unit, but lack of unfaulted good exposures and the scarcity of fossils precludes a full understanding of the stratigraphy and depositional environments. The unit was not studied in detail.

The Lower member consists of siltstone and shale, and locally common volcanogenic sandstone and conglomerate. Based on the fine grained texture and single fossil location, we suggest the member is a marine sequence probably deposited in an outer shelf to slope environment with a dominantly volcanic source region. The local variegated nature of the clastic rocks allows that many depositional facies might be represented in the unit.

Middle member

The Middle member is dominated by laminated siltstone and sandstone (Fig. 44), with minor coarse grained sandstone. Blue-green and greenish-grey siltstone are present in beds of varying thickness up to 3 to 4 m. Grey, fine- to very fine-grained sandstone is interlaminated with the siltstone; laminations are distinct to diffuse. Most sandstone beds show no sedimentary structures but others are ripple laminated. Beds with fine tuff bands are not uncommon. Medium- to coarse-grained sandstones up to 50 cm thick and shale in thin beds a

few centimetres thick occur locally. The coarser sandstones are locally pebbly and calcareous in other places. Clasts are dominated by volcanic rock types.

The Middle member was deposited in a relatively deep marine setting, well below wave base. The minor coarser grained sandstones, and ripple laminated thin sandstones suggest redeposition either in an outer shelf or at the outer margins of a submarine fan. The Middle member grades up into the overlying coarse grained Upper member, interpreted to be a prograding fluvial-influenced, deltaic sequence.

Upper member

The upper member is dominated by lithic sandstone and conglomerate, subordinate laminated siltstone and sandstone, and minor tuff and limestone. The lower 50 m at each location was examined in the most detail and the basal 460 m was examined in less detail 4 km east of Mount Nemaia (Fig. 47). In the lower part of the member, lithic to arkosic-lithic sandstones and volcanic-pebble conglomerates are generally graded in 10 cm to 2 m thick beds, with common mudstone ripup clasts, and minor erosional bases. Thick sandstones in the finer grained beds commonly show ripple and convolute lamination; many are turbidites.

The laminated siltstone and very fine- to fine-grained sandstones appear to be thin, turbidite couplets in the lower part of the section, but these rock types become less laminated upsection, perhaps due to bioturbation. Buff- to white-weathering siltstone layers in the fine grained, laminated facies are possibly tuffaceous in part. Carbonaceous and shelly material is rare to absent low in the member, but increases upsection to where it is common in the coarse grained sandstone beds. A few, minor calcareous sandstones to calc-arenites are present in the upper part of the section. The ~200 m of upper member above the basal ~500 m consists of bundles of sandstone interbedded with 10-30 m shale, laminated siltstone, and limestone sections (Fig. 40).

The Middle and Upper members together are interpreted to be a prograding deltaic sequence; a prograding submarine fan is also a plausible depositional environment. The sequence has the following trends upsection which support this model. (1) The units generally become coarser grained upsection, although there is a fairly abrupt change across a few metres of section between the Middle and Upper members. The upper part of the Lower member continues this grain size trend. (2) Laminated, fine grained rocks low in the sequence include common thin-bedded, turbiditic sandstones and the fine grained rocks become less laminated upsection, probably indicating a greater degree of bioturbation. (3) Mud ripups are common in the sandstone beds throughout the sequence, but carbonaceous debris is distinctly more common high in the section, indicating a greater input from a fluvial source. (4) The presence of fine, shelly material and the more common calcareous cement of the sandstones, high in the sequence, suggest a change into a shallower marine, more biogenic-rich setting.

The upper 200 m of the Upper member is similar to the lower part of the unit. This change upsection may indicate a return to a proximal prodelta.

Biostratigraphy of the Nemaia Formation

Although fossils are sparse and few in number, there is sufficient evidence to suggest that the Nemaia Formation is mainly or entirely correlative with the Last Creek Formation (Fig. 7, 38). Preliminary data suggest that parts of the Nemaia Formation were deposited in deeper water and some parts accumulated closer to a volcanic source than the Last Creek Formation. Because of the scarcity of fossil evidence, the completeness of the sequence can only be interpreted from the greater thickness and lack of evidence of stratigraphic interruptions.

The Lower member is not exposed in any one section. The oldest fauna is of Early Sinemurian and includes the ammonite *Arnioceras* cf. *A. kwakiutlanus* (Crickmay) and a few indeterminate bivalves. This probably correlates with the basal beds of the Little Paradise Member which also carries species of *Arnioceras*. Unfossiliferous shale-siltstone beds underlie this fauna and many metres lower a limestone cobble conglomerate was noted; this might suggest Triassic rocks but there is no fossil evidence.

An Early Pliensbachian fauna is found in a thin conglomeratic sequence near the north end of Tatlayoko Lake (see above). There a fauna includes the ammonite *Dubariceras freboldi* Dommergues, Mouterde and Rivas, the bivalve *Weyla* and a few indeterminate bivalves. It is of interest to note that this fauna in a conglomeratic sequence is precisely correlative with the only conglomerate in the Little Paradise Member of the Last Creek Formation. As conglomerates are essentially unknown in the Lower Jurassic beds above the Sinemurian in Last Creek Formation, it is reasonable to suggest the possibility of a widespread erosional interval or for a drop in sea level in late Early Pliensbachian time (see Smith and Tipper, 1986; Smith et al., 1988); the latter is preferred.

No Toarcian fossils have been found in the lower member of the Nemaia Formation but an excessively thick sequence of unfossiliferous shale and siltstone underlies Aalenian beds in the Nemaia Range and could represent much of the Toarcian.

On Huckleberry Mountain, at the south end, Aalenian ammonites were found in laminated black shales near or at the top of the Lower member. A few fragments were found in 1967 and were reported on by Poulton and Tipper (1991).

In 1990 a more extensive collection was made from several localities but the stratigraphic relations were not decipherable. The fauna was identified by T.P. Poulton who determined the following:

Planamatoceras sp.
Tmetoceras sp.
Planamatoceras(?) (Pseudaptetoceras?) sp.
Erycitoides sp.

The Middle member of the formation in the type section yielded only a sonniniid(?) ammonite but elsewhere in the area in isolated outcrops several forms were found that probably represent the fauna of this interval (earliest Early Bajocian?) and include *Sonninia(?)* sp. aff. *bifurcata* Imlay and other indeterminate forms. Bivalves are rare.

The fauna of the Upper Member is dominated by Early Bajocian stephanoceratids but specimens are commonly fragmented. Several forms have been recognized:

Stephanoceras sp. cf. *skidegatense* (Whiteaves)
Zemistephanus sp. cf. *richardsoni* (Whiteaves)
Pseudolioceras(?) crickmayi (Frebold, Tipper and Coates)
Chondroceras sp.
Stephanoceras sp.
Oppelia(?) sp. aff. *stantoni* Imlay
Oppelia sp.
Stemmatoceras sp.
Hebetoxyites(?) sp.

One collection with *Leptosphinctes(?)* sp. and small sphaeroceratid(?) ammonites suggests a Late Bajocian(?) age for the youngest fauna of the formation.

The Nemaia Formation apparently ranges in age from mid-Early Sinemurian to possibly Late Bajocian with no major unconformity except possibly in late Early Pliensbachian. The formation apparently rests unconformably(?) on Late Triassic rocks of the Tyaughton and Hurley formations and is overlain unconformably (disconformably?) by (?)[Late Bathonian]-Callovian sediments.

THE TRIASSIC-JURASSIC BOUNDARY

The Triassic-Jurassic (T-J) boundary in the Castle Pass area corresponds to the contact between the Tyaughton Formation and the Last Creek Formation (see cover photograph). The upper green clastics member of the Tyaughton Formation has yielded ammonoids identified as *Choristoceras crickmayi* which, according to Tozer (1967), indicate the youngest North American zone, the Crickmayi Zone of the Upper Norian (=Rhaetian of Europe). The conodont *Misikella posthernsteini* Kozur and Mock was also found in this member and marks the highest level of identifiable conodonts; in Queen Charlotte Islands this conodont occurs a few metres below the Triassic-Jurassic boundary defined by radiolarians (Tipper et al., 1994). On the basis of this evidence, there is a strong probability that the Tyaughton Formation ranged to the end of the Triassic, possibly even into the Jurassic. Hettangian faunas of the Castle Pass Member of the Last Creek Formation are found in the basal conglomeratic parts of that member. In Queen Charlotte Islands a complete or nearly complete Hettangian sequence is documented by fossils, mainly ammonites, and this sequence rests directly on a fossiliferous sequence of youngest Triassic rocks; a continuum across the Triassic-Jurassic boundary is postulated (Tipper et al., 1994). The Hettangian faunas of the Castle Pass Member near Castle Peak are closely similar to those in the upper one-third of the Hettangian section of Queen Charlotte Islands (Tipper and Guex, 1994). It is reasonable to suggest that these are "Late" Hettangian in age and that "Middle and Lower" Hettangian strata are not recognized in the Castle Peak area although Frebold (1967) dated a form as early Hettangian. Frebold described a form as *Psiloceras* ex aff. *P. planorbis* (Sowerby) thus indicating an Early Hettangian age. A re-examination of this specimen by Guex (1980) and Tozer (1982b) led to the conclusion that identification was erroneous and the specimen was a Triassic form, possibly a *Rhacophyllites*.

In places the Castle Pass Member of the Last Creek Formation rests directly on *Cassianella* beds member of the Tyaughton Formation. The upper green clastics member was either eroded before deposition of the Hettangian sediments or, in places, the upper green clastics member was never deposited; erosion is the preferred explanation.

The Triassic-Jurassic boundary in the Castle Peak area is apparently marked by an erosional interval that may have lasted for the early two-thirds of the Hettangian. The basal Jurassic pebble conglomerate contains clasts derived from the Triassic upper green clastics member as well as clasts of lithologies that are not represented as members of the Tyaughton Formation, such as volcanic rocks and rare plutonic rocks. These could be explained as being eroded from older Triassic formations such as the Cadwallader Group thus suggesting a deep and widespread erosion of Triassic and older(?) groups or formations. The simplest explanation is the reworking of Tyaughton Formation conglomeratic members that do contain volcanic and rare plutonic clasts and there is no need to call upon widespread deep erosion. Furthermore there is no compelling evidence to suggest in this area an angular discordance at Triassic-Jurassic boundary or a period of strong deformation. A small vertical uplift or sea- level drop and erosion is all that is required to explain the relationships.

In British Columbia the only proven Hettangian strata are in the Queen Charlotte Islands and in the Tyaughton Creek area. The direction of transport in the Tyaughton Formation and in the Hettangian beds of the Last Creek Formation is to the west and southwest. It is interesting to speculate that the Tyaughton-Last Creek basin is the eastern extension of the Triassic-Jurassic basin of Queen Charlotte Islands receiving sediment from Cache Creek, Quesnellia, the craton, or other eastern terranes.

PROVENANCE

Tyaughton Formation and Last Creek Formation

The clastic rocks of the Tyaughton Formation and Last Creek Formation define two distinct petrofacies, both derived from an inactive magmatic arc. The oldest members (basal red conglomerate and lower green clastics – *Cassianella* beds) of the Tyaughton Formation comprise a mixed plutonic- and volcanic-derived petrofacies with a minor metamorphic component. Sandstone composition is subequally quartz, feldspar, and rock fragments of both plutonic and volcanic origin (Fig. 48A) and the rocks classify as relatively quartz-rich, lithic arenites and arkosic arenites (classification of Pettijohn et al., 1973). Conglomerates of the mixed petrofacies are dominated by either intermediate to felsic volcanic clasts or

limestone clasts, with a minor plutonic component. The composition of conglomerate clasts is similar to the slightly older conglomerates of the Hurley Formation (Rusmore, 1985, 1987) and, together with biostratigraphic data, strongly suggests that the older units of the Tyaughton Formation were derived from the same region as the Hurley Formation, or, in part, from erosion of the Hurley Formation itself.

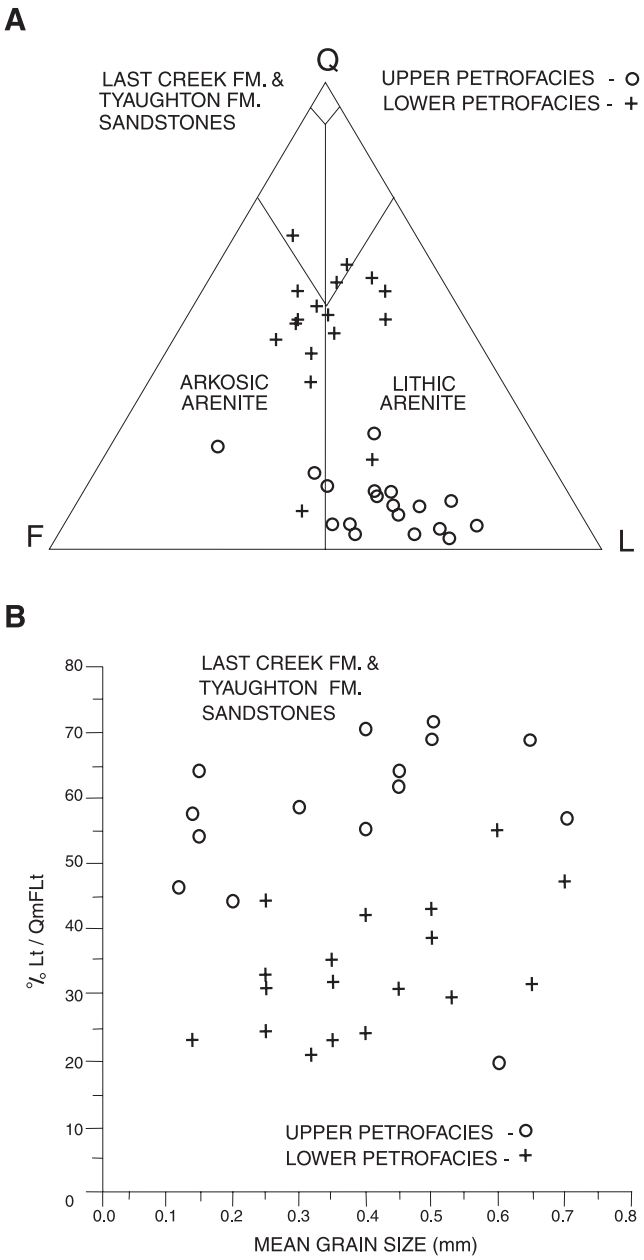


Figure 48. A) *Sandstone composition of samples from two petrofacies of the Tyaughton Formation and Last Creek Formation plotted on a quartz (Q), feldspar (F), lithic fragments (L) diagram with composition fields from Pettijohn et al. (1973).* ***B)*** *Plot of mean grain size versus per cent of lithic fragments in sandstones of the Tyaughton and Last Creek formations, which demonstrates that there is only a weak linear relation between the two parameters.*

The second petrofacies is a volcanic-derived facies made up of the upper green clastics of the Tyaughton Formation and the overlying Castle Pass Member of the Last Creek Formation. The sandstones in this facies are quartz-poor, lithic arenites (Fig. 48A). Conglomerates contain mainly volcanic clasts, minor plutonic clasts, and rare limestone. A minor, but ubiquitous, plutonic signature is present in the sandstones. A change to a mixed plutonic and volcanic source area occurs in the upper part of the Castle Pass Member.

Sandstone petrography – methodology

Samples of sandstones were collected throughout the exposure of the Tyaughton Formation and Last Creek Formation both north of Tyaughton Creek and east of Spruce Lake (Fig. 4) (Umhoefer et al., 1988). Thin sections were made from 133 samples and 37 of the least-altered sections were point counted. Alteration and/or replacement of lithic and feldspar grains by secondary minerals is ubiquitous, but in nearly all of the sections that were counted, the framework grain boundaries were readily recognized and identification of the original grain type was possible. Some limitations due to alteration are discussed below. Using Dickinson's (1970) terminology, "pseudomatrix," or the breakdown of lithic grains to form matrix-like material is common in many sandstones in the Tyaughton and Last Creek formations, but enough samples lack pseudomatrix to successfully investigate sandstone composition by means of point counting.

The Gazzi-Dickinson point counting method (Dickinson, 1970; Ingersoll, 1983) was employed so that sand-sized components in polycrystalline rock fragments were counted as individual grains. Grains smaller than 0.625 mm were considered matrix. Over 300 framework grains were counted in each section at a spacing of 0.5 mm. Grain parameters follow Ingersoll and Suczek (1979).

Calcareous sandstones and sandy calc-arenites are present in a number of the units. These calcareous samples were examined in thin section, and a few of them were point counted. As would be expected in an analysis with a closed system that totals 100%, these samples were anomalous when compared to samples in the same unit without calcareous framework grains (Dickinson, 1970); therefore they were not included in the following analysis.

About one-third of the thin sections were stained for plagioclase and K-feldspar. The staining hindered identification of lithic grains and did not always help identification of plagioclase and K-feldspar because of the common alteration of feldspars by sericitization, vacuolization, and calcite and clay replacement. No K-feldspar was identified by staining, and the resolution of plagioclase versus K-feldspar is not considered reliable, except where obvious albite twinning is present. It is reasonable to expect that K-feldspar would be albitized during diagenesis, because the Tyaughton Formation was probably buried at least 4 km by the Relay Mountain and Taylor Creek groups (Umhoefer, 1989; Garver, 1992), and recent work has shown that albitization of K-feldspar is common below about 3.5 km (Saigal et al., 1988). Despite this limitation, feldspar was readily distinguished from quartz because quartz is distinctly less altered than feldspar.

The Gazzi-Dickinson method is not as sensitive as other methods to variations of composition with changes in grain size in medium- to coarse-grained sandstones (Ingersoll et al., 1984). In some rock units, fine grained sandstones do show significant compositional variations when compared to coarser sandstones. The thin sections used in this study include samples which are fine-, medium-, and coarse-grained; therefore the effect of composition on grain size was investigated. Figure 48B demonstrates that there is no systematic variation away from the average per cent of lithic grains when plotted against grain size; therefore, all samples were included in the analysis.

The Gazzi-Dickinson method was explicitly developed to determine the lithotectonic character of the source region of a sandstone, which was the major objective of this petrographic study. The method has been successfully tested in modern basins (Ingersoll and Suczek, 1979; Valloni and Maynard, 1981; Valloni and Mezzadri, 1984; Suczek and Ingersoll, 1985; Valloni, 1985; Thornberg and Kulm, 1987; Girty et al., 1988) and in ancient sequences where the tectonic setting of the basin is known (Dickinson et al., 1979, 1983; Ingersoll, 1983; and many others). Similar compositional patterns emerge from most studies and the general applicability of relating sandstone modes to tectonic setting of the basin seems firmly grounded in empirical observation. This has led to common use of provenance-field diagrams which assume that particular QFL or QmFLt compositions relate to a unique tectonic setting (Dickinson and Suzcek, 1979; Dickinson et al., 1983). These diagrams are used below although we fully realize that a number of workers have demonstrated limits to the simple approach of relating provenance-field diagrams to tectonic setting (Velbel, 1985; Saccani, 1987; Girty et al., 1988).

One limitation in the use of QFL diagrams is relevant to the current study. Girty et al. (1988) sampled modern river sands from the western flanks of the Peninsular Ranges batholith of southern California and Baja California. Sampling was restricted to areas that drain only Mesozoic rocks – the Cretaceous batholith and older metasedimentary and metavolcanic country rocks; the sands generally fall in the arc field on provenance diagrams (Girty et al., 1988). Two exceptions were found: sands from purely batholithic outcrop areas sometimes plot near the Q-F line on QFL plots indicating a basement uplift setting, and sands from dominantly metasedimentary areas plot in the recycled orogen field. Girty et al. (1988) warned that studies conducted in ancient rocks that sample only a limited geographical area may fortuitously sample an area with an "incorrect" signature. The Tyaughton Formation is limited in areal exposure to the relatively small area near Tyaughton Creek (Fig. 4). Despite this limitation, the data are consistent with stratigraphic and sedimentological arguments discussed throughout this report which point to a depositional setting at the edge of an inactive magmatic arc.

Lower Tyaughton Formation petrofacies

The oldest petrofacies includes the basal red conglomerate and lower green clastics to *Cassianella* beds members (the latter hereafter included with the lower green clastics). All point count data are listed in Table 1 and plotted on various plots in Figures 49 and 50. The point count data plot in the centre of QFL and QmFLt diagrams (Fig. 49). The basal red conglomerate member has a dominant subfacies that plots in the recycled orogen field on the QFL plot of Dickinson et al. (1983) and in the mixed arc and recycled field on the QmFLt plot; two other points lie in the transitional arc field. The two samples more rich in volcanic detritus are spatially separated from the others, outcropping to the east of Spruce Lake, and the others are from the main outcrop belt north of Tyaughton

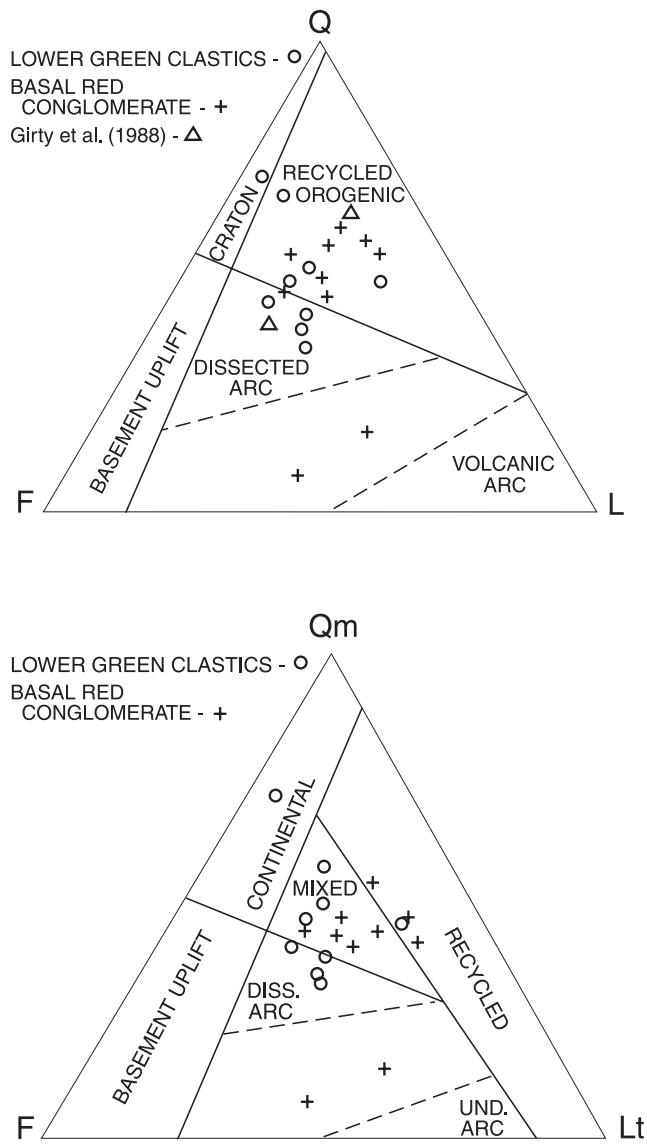


Figure 49. *Provenance plots of lower petrofacies of the Tyaughton Formation. Provenance fields after Dickinson et al. (1983). Ternary diagram parameters: Q, total quartz; L, 'unstable' aphanitic lithic fragments; Qm, monocrystalline quartz; F, feldspar; Lt, lithic grains including polycrystalline quartz. Data in Table 1. See text for discussion of Girty et al. (1988) data; diss. arc = dissected arc; und. arc = undissected arc.*

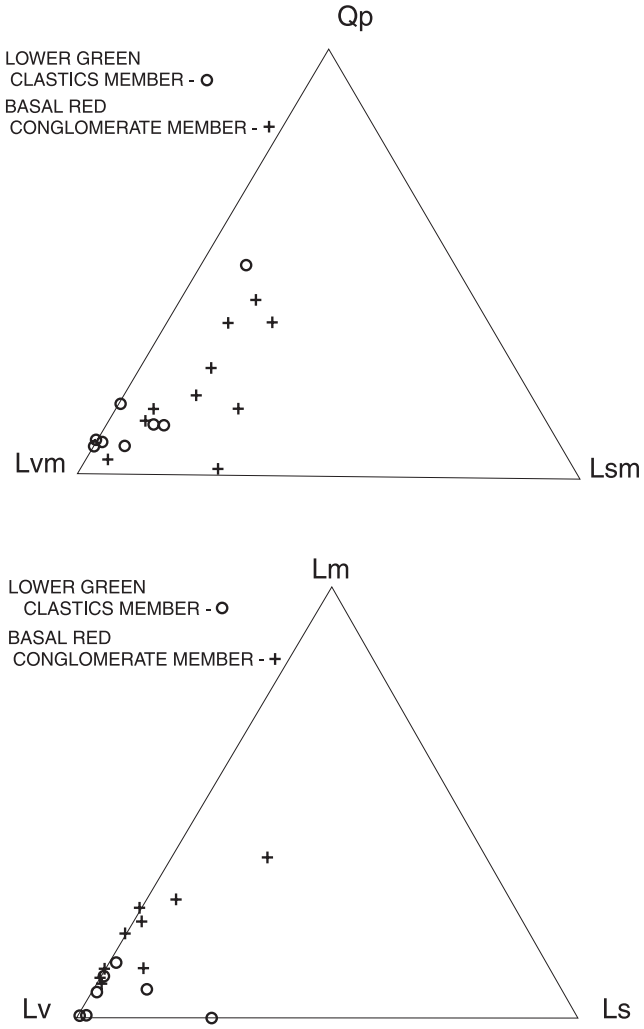


Figure 50. *Provenance plots of lithic clasts of lower petrofacies, Tyaughton Formation. Ternary diagram parameters: Qp, polycrystalline quartz; Lvm, volcanic-hypabyssal and metavolcanic aphanitic lithic grains; Lsm, sedimentary and metasedimentary aphanitic lithic grains; Lm, metamorphic aphanitic lithic grains; Lv, volcanic aphanitic lithic grains; Ls, sedimentary aphanitic lithic grains. Data in Table 1.*

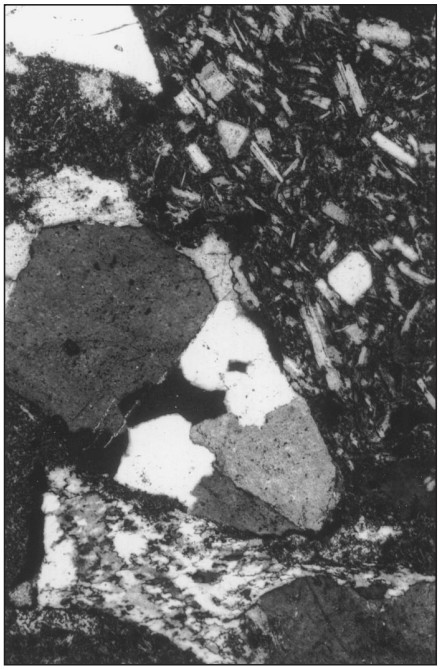
Creek (Fig. 4) (Umhoefer et al., 1988). The sandstones of the lower green clastics member bridge the dissected arc and recycled orogen fields on the QFL plot and the arc and mixed fields on the QMFLt plot. The lithic fraction varies from nearly all volcanic lithics (Lvm) to 50% polycrystalline quartz (Qp) in the basal red conglomerate member and up to 15% Qp in the lower green clastics member, both typical of basins adjacent to volcanic arcs (Dickinson and Suzcek, 1979; Ingersoll and Suczek, 1979). A small metamorphic (5-10%) component makes up the basal red conglomerate member; this becomes smaller in the lower green clastics member.

Three significant patterns are inferred in the ternary plots for the mixed plutonic and volcanic petrofacies. First, there is a definite magmatic arc signal, but mixing with a "recycled orogen" component needs explanation. Second, there is an unequivocal metamorphic signal. Third, the trend from the older basal red conglomerate member to the lower green clastics member shows development of a more dominant arc constituent and loss of most of the metamorphic constituent.

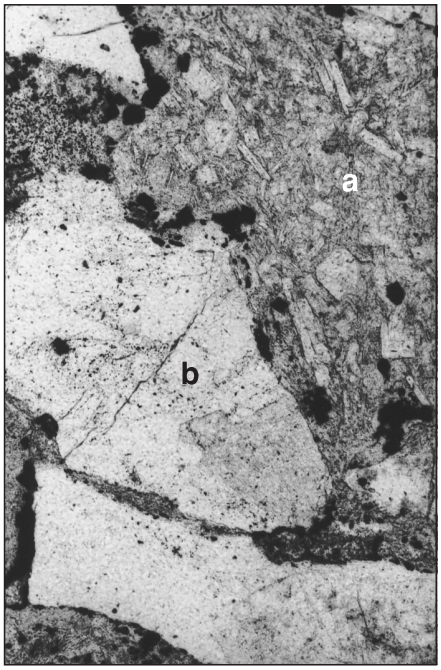
Individual characteristics of the sandstones support a mixed plutonic and volcanic source dominated by plutonic rocks, that is a setting on the fringe of a dissected magmatic arc. Twenty to fifty per cent of the monocrystalline quartz grains were sand-sized grains within a polycrystalline aggregate of plutonic origin – either 2 or 3 grains of quartz or quartz+feldspar (Fig. 51A, B, D, E). In most samples up to 50% of the feldspar grains were also part of plutonic fragments. Microplutonic grains represent 5 to 15% of the grains in all samples (Fig. 51C). These grains are quartz+plagioclase aggregates interpreted to be hypabyssal volcanics. A few grains in each sample are graphic granite and myrmekite, both derived from granitoid source rocks. The former is the intergrowth of quartz and K-feldspar and is most common in acid granites, and the latter is an intergrowth of plagioclase and quartz in the presence of K-feldspar. These observations are important because confirmation of K-feldspar was only rarely achieved, and stained samples confirm that it is uncommon. The range of intrusive rocks in the source area was probably from granite to tonalite. Finally, although volcanic fragments are subordinate to plutonic-derived clasts, volcanics make up 10 to 25% of the clasts (Fig. 51A, B, D, E), and no doubt some of the monolithic quartz and plagioclase is volcanic in origin (Fig. 51D, E). Some of the twinned plagioclase grains are elongate, nearly euhedral laths of volcanic origin. The mix of felsic, intermediate, and mafic volcanic clasts varies from sample to sample, but the entire range is represented in all samples. Textures described as microgranular, microlitic, and lathwork represent the range of volcanic compositions.

The metamorphic component is evident in both lithic grains and quartz. Metamorphic lithic grains are 1-2% (of total grains) quartz-mica tectonites (Fig. 51F) and 2-5% foliated quartz tectonites. These grains are most likely to have originated in a high-grade metamorphic terrane. The polycrystalline quartz is not chert, but is entirely of plutonic or metamorphic origin. Some of the felsic volcanics are slightly recrystallized and may be metavolcanic in origin (Fig. 51A, B). Finally, undulose quartz dominates over quartz with nearly straight extinction. Undulose grains could be caused by low levels of strain within the sandstones as evidenced by common pressure solution and grain-to-grain-contact features. Most sandstones have 10-15% true matrix, which can retard widespread formation of undulosity (Basu et al., 1975). Another explanation for the common undulose quartz is that it originated in low- to medium-grade metamorphic rocks. If these rocks are used for quartz analysis following Basu et al. (1975), then the sandstones of the lower petrofacies plot in the low-grade metamorphic field (Fig. 52). This interpretation is made with caution. Many of the samples are not strictly appropriate for quartz analysis because they are medium- to coarse-grained instead of only medium grained (particularly in the basal red conglomerate), and a few samples have a low per cent matrix which can facilitate undulose quartz formation as discussed above. In summary, in the extreme interpretation,

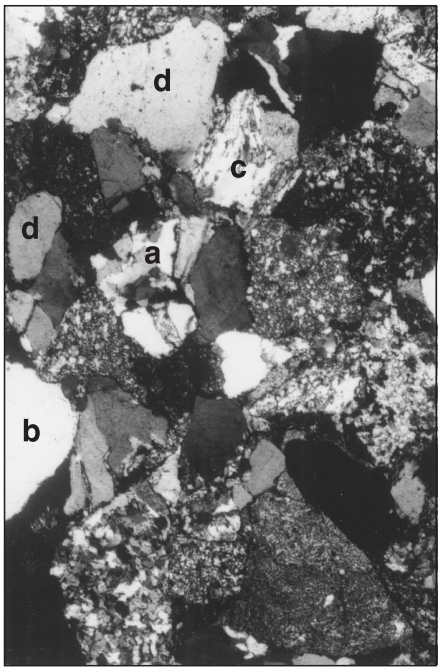
Figure 51. *Photomicrographs of sandstones of the Tyaughton, Last Creek, and Nemaia formations.*



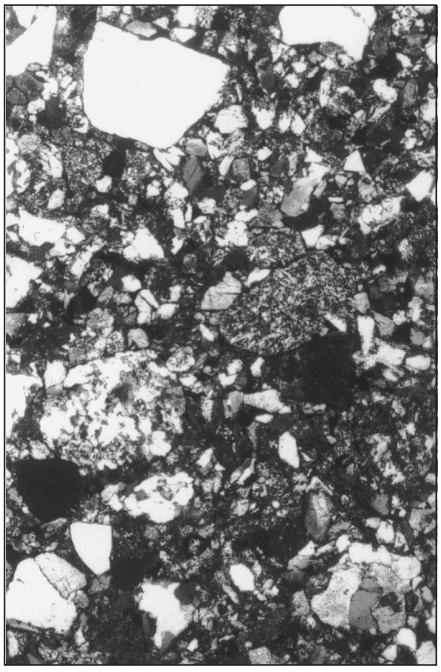
A) *Sample UT85-52 from basal red conglomerate member, Tyaughton Formation; (a) typical andesite with plagioclase laths in upper right and (b) plutonic rock fragment with plagioclase and quartz in left centre (see Fig. 51B for labels); photo in polarized light and 1.2 mm from top to bottom.*



B) *Same sample and view as Figure 51A, but with plane light.*

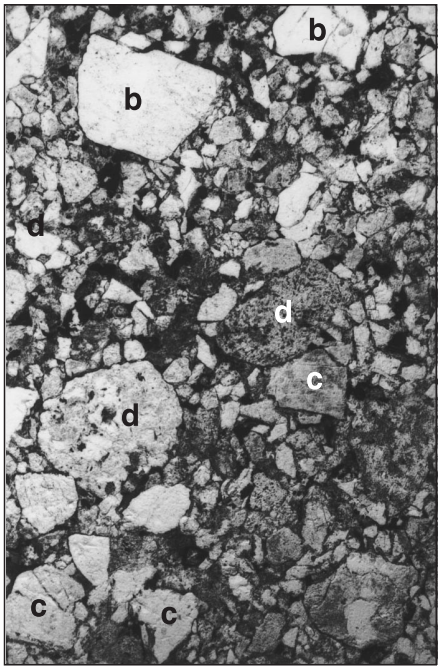


C) *Sample UT85-96 from basal red conglomerate member, Tyaughton Formation; (a) micro-plutonic rock fragment; (b) quartz; (c) metamorphic fragment or strained micro-plutonic fragment; most of the other grains are either (d) plagioclase or volcanic rock fragments; polarized light and 3 mm from top to bottom.*



D) *Sample UT85-85-23 from lower green clastics member, Tyaughton Formation; large grains on left and in middle are quartz-mica tectonites, otherwise rare to nonexistent in the Tyaughton Formation; polarized light and 1.2 mm top to bottom.*

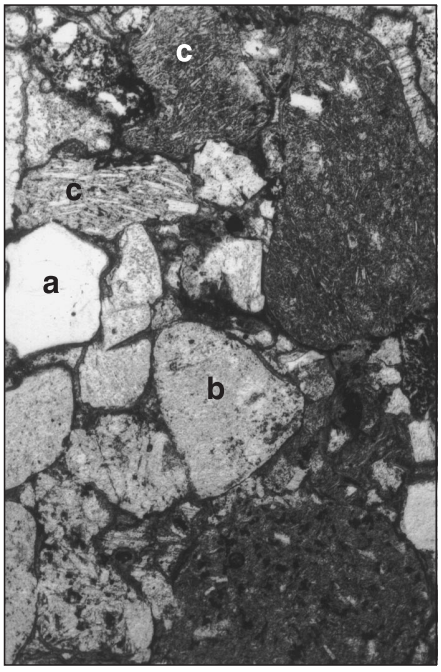
Figure 51. *(cont.)*



E) *Sample UT85-176B from lower green clastics member, Tyaughton Formation; note poor sorting and common angular to subangular grains; polarized light and 3 mm from top to bottom.*

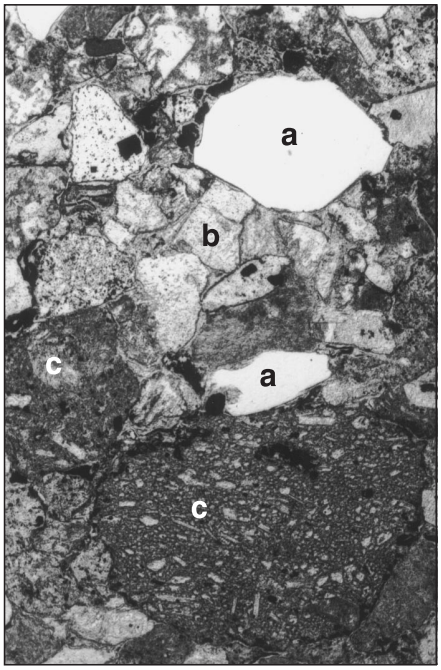
F) *Same sample and view as Figure 51E above; (a) uncommon micro-plutonic grain; and (b) quartz; common (c) plagioclase and (d) volcanic rock fragments; plane light and 3 mm from top to bottom.*

G) *Sample UT85-258F from upper green clastics member, Tyaughton Formation; polarized light and 3 mm from top to bottom.*

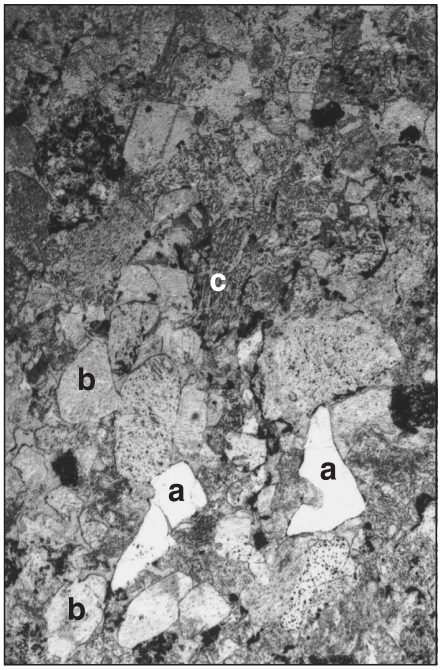


H) *Same sample, view, and scale as Figure 51G above; (a) uncommon embayed (volcanic) quartz; most of the other grains are (b) felsic volcanic and andesite, some (c) with flow-aligned micro-plagioclase; plane light.*

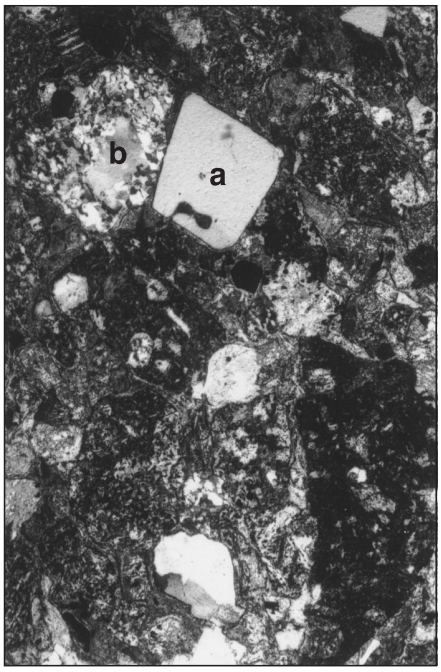
Figure 51. *(cont.)*



I) *Sample UT85-317-16 from Castle Pass Member, Last Creek Formation; (a) angular and embayed quartz; (b) common plagioclase and (c) volcanic rock fragments (large grain at bottom); plane light and 3 mm top to bottom.*



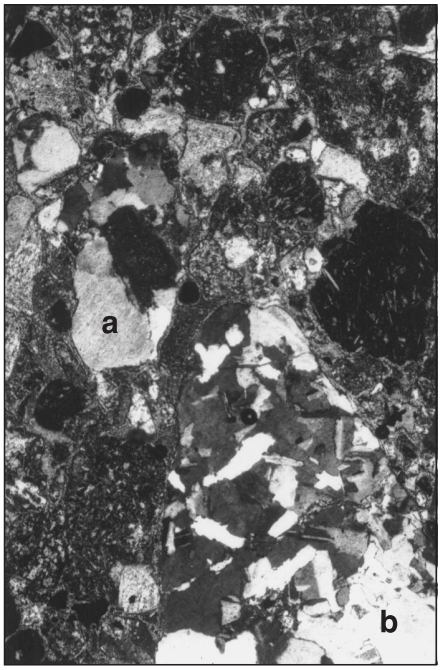
J) *Sample UT85-316-10 from Castle Pass member, Last Creek Formation; (a) embayed and angular (volcanic) quartz, most other grains are (b) plagioclase and (c) volcanic rock fragments; plane light and 3 mm from top to bottom.*



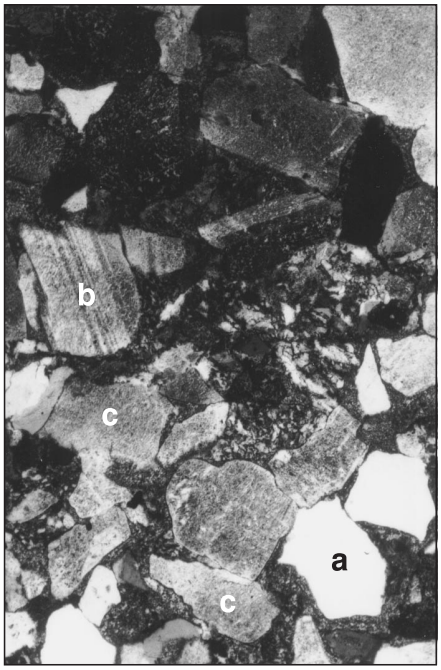
K) *Sample UT85-314-11 from Castle Pass member, Last Creek Formation; (a) embayed (volcanic) quartz; (b) strained polycrystalline quartz; most other grains are plagioclase and volcanic rock fragments; polarized light and 3 mm top to bottom.*

L) *Same sample, view, and scale as Figure 51K; plane light.*

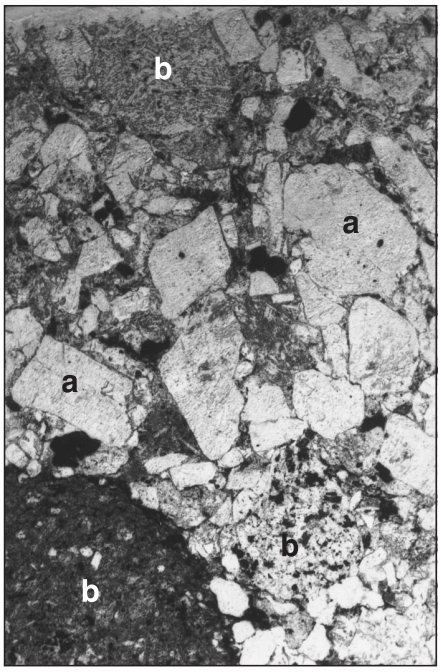
Figure 51. *(cont.)*



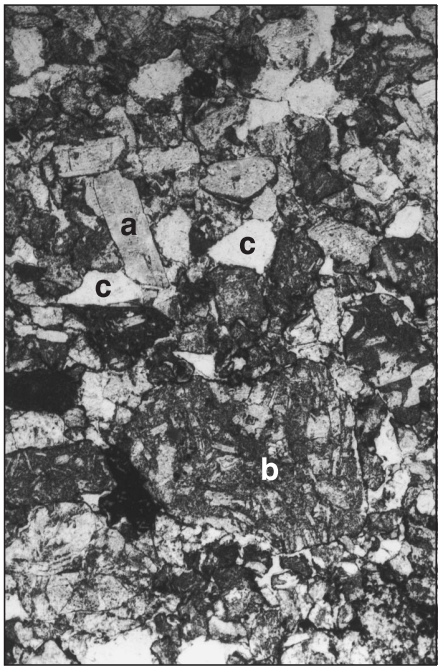
M) *Sample UT85-314-11 from Castle Pass Member, Last Creek Formation; (a) plutonic rock fragments and dark grains in lower left, central right, and central top are andesite to basaltic andesites; polarized light and 3 mm top to bottom.*



N) *Sample UT85-185 from Castle Pass member, Last Creek Formation; (a) quartz and common (b) twinned and (c) untwinned plagioclase in a plagioclase-rich arkose; polarized light and 3 mm top to bottom.*



O) *Sample U90-LC-9A from Nemaia Formation; typical sandstone with poor sorting and angular to subrounded (a) plagioclase and (b) volcanic rock fragments; plane light and 3 mm top to bottom.*



P) *Sample U90-LC-16-4 from Nemaia Formation; rich in (a) plagioclase and (b) volcanic rock fragments and minor (c) quartz; plane light and 3 mm top to bottom.*

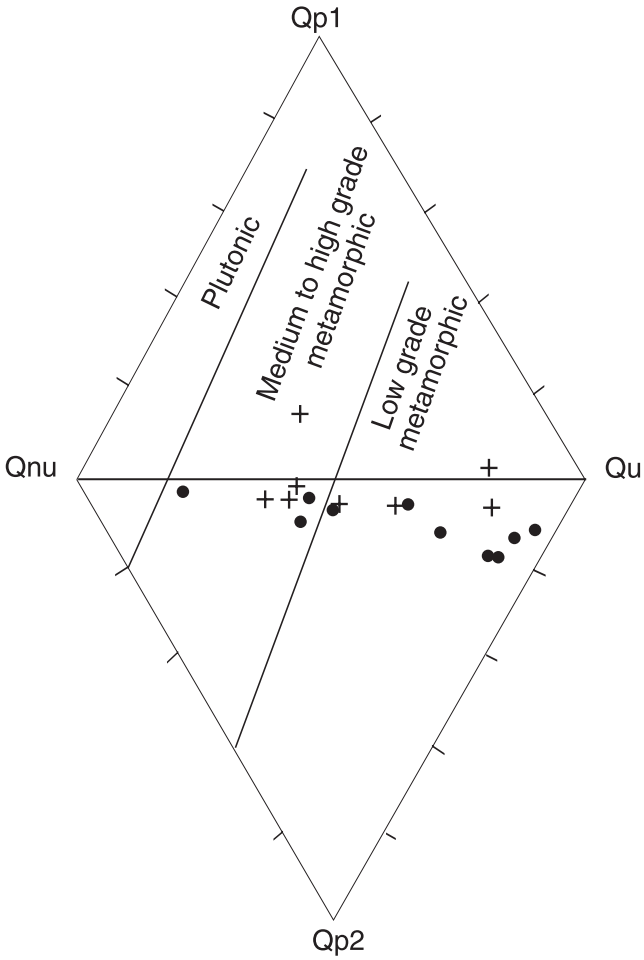


Figure 52. *Quartz-provenance diagram of lower petrofacies, Tyaughton Formation. Diagram after Basu et al. (1975). +'s = lower green clastics samples; circles = basal red conglomerate samples. Parameters: Qpl, >75% of Qp grains is made of grains with <4 crystals per grain; Qp2, >25% of Qp grains is made of grains with <4 crystals per grain; Qnu, nonundulatory monocrystalline quartz; Qu, undulatory monocrystalline quartz. Undulation based on >5° rotation of extinction of grain.*

where all undulose quartz is considered metamorphic in origin, 20-50% of the grains in the basal red conglomerate were derived from metamorphic rocks and 15-30% in the lower green clastics member. The minimum extreme for metamorphic derivation of clasts is 2-10% in the basal red conglomerate unit and 0-3% in the lower green clastics. The metamorphic signal is the best explanation for the data plotting in both the arc and recycled orogen fields on the ternary diagrams (Fig. 49, 50).

The changes in sandstone petrography from the basal red conglomerate member(particularly when the dominant subfacies is considered alone) to lower green clastics member are small, but include (a) a decrease in polycrystalline quartz and increase in feldspar in the lower green clastics member, (b) a decrease in metamorphic clasts in the lower green clastics member, and (c) a change from angular/subangular grains in the basal red conglomerate member to subangular/subrounded grains in the lower green clastics member. All of these trends indicate an increase in maturity of sediment in the lower petrofacies. The greater rounding upsection is probably the result of the different environments of deposition of the units. The basal red conglomerate member comprises deposits of alluvial-fan debris flows and proximal braided-streams, whereas the lower green clastics member is deposits of more distal braided streams that eventually grade up into nearshore deposits. The small petrofacies changes probably represent either a degradation of more unstable clasts in the more distal unit, or a decrease in the availability of metamorphic fragments in the source region. Alternatively, the greater susceptibility to weathering in the soil profile for polycrystalline versus monocrystalline quartz (Basu, 1985) could explain the changes.

The small changes in the lower petrofacies may best be explained from a component of recycling of the basal conglomerate. An unconformity at the base of the intervening limestone conglomerate (Fig. 9) and the lithological character of the limestone conglomerate demonstrate uplift of the basin shortly after deposition of the massive limestone unit, that is between deposition of the basal red conglomerate and lower green clastics members. The lower green clastics (sandstones) may represent erosion of the uplifted basal red conglomerate and mixing of red conglomerate detritus with the background magmatic arc source. Chemical weathering of the basal red conglomerate would preferentially eliminate the Qp from the sand population compared to the Qm, and produce the petrofacies trends. Reworking of the basal red conglomerate also would produce the more rounded grains in the sandstones of the lower green clastics member.

A better constrained analogy for the petrography of the lower Tyaughton petrofacies may be the study of modern sands derived from the Peninsular Ranges of Baja California and southern California, briefly summarized above (Girty et al., 1988). The metasedimentary bedrock in that study produced relatively quartz-rich, fluvial sands which plot in the same part of the recycled orogen field as the more quartz-rich sandstones in the lower Tyaughton petrofacies (Fig. 49). A hypothetical mix of the metasedimentary, plutonic, and metavolcanic components in the Peninsular Ranges study produces a sand composition close to the lower Tyaughton petrofacies (Fig. 49).

Upper Tyaughton Formation – lower Last Creek Formation petrofacies

The younger volcanic petrofacies consists of the upper green clastics member of the Tyaughton Formation and the immediately overlying coarse clastic part (Castle Pass Member) of the Last Creek Formation. We combine samples from two members into a single petrofacies to emphasize the similarities in petrography; there are small changes across the formation boundary that are summarized below. The change from the lower to upper petrofacies occurs over a very short stratigraphic interval, and in a very short time, in the uppermost *Cassianella* beds member. Only two sandstones in the

Cassianella beds yielded thin sections appropriate for point counting; those are similar to the lower green clastic member, in which they were included. Samples at the transition from *Cassianella* beds to upper green clastics member show the rapid change to the upper facies over a few metres of section.

All point count data for the upper petrofacies are listed in Table 1 and shown on ternary diagrams in Figures 53 and 54. The data for the upper petrofacies are very distinct from the lower petrofacies, a point that will be discussed below. The upper facies plots in the undissected arc to transitional arc fields on QFL and QmFLt diagrams, typical of many classic arc-related sequences (Dickinson et al., 1979, 1983). Lithic plots also show the dominance of volcanic detritus. The QpLvmLsm plot shows nearly all volcanic lithics (Fig. 51G, H, I, J) and a few per cent polycrystalline quartz (Qp). Most, or all, of the Qp is plutonic in origin. Likewise, nearly all the LmLvLs grains are volcanic lithics; a few samples contain a small per cent sedimentary lithics. In addition, most quartz grains are angular and show a texture indicating a volcanic origin (Fig. 51G, H, I, J, K, L).

The petrographic data of the upper facies clearly show that the upper green clastics member and Last Creek Formation sandstones were derived from the erosion of a relatively undissected (and young?) volcanic arc (Fig. 53). The data are

Table 1. Percentages for petrographic parameters, Last Creek Formation and Tyaughton Formation sandstones. Petrographic parameters are defined in Figures 49 and 50.

UNIT	SAMPLE	Qt	Qm	F	L	Lt	M	Qp	Lvm	Lsm	Lm	Lv	Ls
LAST CREEK FORMATION:													
LC	UT85-185	21	21	59	20	20	0	3	95	2	0	98	2
LC	UT85-314-3	2	2	26	72	72	0	0	97	3	0	97	3
LC	UT85-314-6B	5	3	46	49	51	0	4	96	0	0	100	0
LC	UT85-314-7	12	10	36	53	54	0	2	97	1	1	99	0
LC	UT85-314-11	9	9	34	58	58	1	0	98	2	0	98	2
LC	UT85-314-13	5	5	43	52	52	0	0	95	5	0	95	5
LC	UT85-316-3	16	14	44	40	42	0	6	94	0	0	100	0
LC	UT85-316-10	3	3	32	65	66	0	1	96	4	0	96	4
LC	UT85-317-14	24	21	29	47	50	1	6	33	61	2	35	63
LC	UT85-317-16	10	9	22	68	69	0	1	93	6	0	94	6
TYAUGHTON FORMATION:													
UG	U86-316-A	3	3	43	54	54	0	0	99	1	0	99	1
UG	UT85-140B	7	7	33	60	60	0	1	97	2	0	98	2
UG	UT85-15	9	7	29	63	64	1	2	67	31	0	68	32
UG	UT85-258F	4	4	27	69	69	0	0	100	0	0	100	0
UG	UT85-295B	12	8	32	56	60	1	7	81	12	0	87	13
UG	UT85-317-3A	11	10	35	54	55	1	1	95	4	0	96	4
UG	UT85-317-6	5	4	20	75	76	2	1	69	30	0	69	31
UG	UT85-317-10	13	13	43	44	44	1	0	97	3	0	97	3
CS	UT85-254	44	40	37	19	23	2	16	84	0	0	100	0
CS	UT85-82-35	48	46	31	21	23	1	7	93	0	0	100	0
LG	UT85-82-20	66	56	22	11	21	0	48	43	9	6	83	11
LG	UT85-82-26	35	33	35	30	32	1	7	92	1	0	99	1
LG	UT85-82-29	41	38	32	27	30	0	11	80	9	9	90	1
LG	UT85-85-23	48	45	15	37	40	0	6	87	6	5	93	2
LG	UT85-176B	38	34	34	28	31	1	11	77	11	12	86	2
LG	UT85-223	51	49	27	23	24	0	7	92	1	0	99	1
LG	UT85-108C	70	70	25	5	6	0	6	94	0	0	100	0
BR	UT85-41	8	8	50	42	42	2	1	73	27	0	73	27
BR	UT85-52	17	15	33	50	52	0	3	93	4	11	89	0
BR	UT85-84A	54	46	12	34	41	0	18	69	14	22	76	2
BR	UT85-85-5	60	53	16	24	31	0	24	62	14	19	81	0
BR	UT85-85-11	54	42	28	18	30	0	40	45	15	25	75	0
BR	UT85-85-13	57	41	13	30	45	0	35	45	21	37	44	19
BR	UT85-96	56	43	20	24	37	0	35	53	12	11	81	8
BR	UT85-167B	45	40	26	29	34	1	15	78	7	7	91	2
BR	UT85-200	49	46	25	26	29	0	12	82	7	9	91	0
BR	UT85-216	47	43	33	21	24	0	15	61	24	27	67	6

Qt – total quartz
F – feldspar
Lt – lithic grains including polycrystalline quartz
Qp – polycrystalline quartz
Lsm – sedimentary and metasedimentary aphanitic lithic grains
Lv – volcanic aphanitic lithic grains
LC – Last Creek Formation
CS – *Cassianella* beds
BR – basal red conglomerate
Qm – monocrystalline quartz
L – unstable aphanitic lithic fragments
M – matrix
Lvm – volcanic-hypabyssal and metavolcanic aphanitic lithic grains
Lm – metamorphic aphanitic lithic grains
Ls – sedimentary aphanitic lithic grains
UG – upper green clastics
LG – lower green clastics

indistinguishable from sands in Neogene basins along the central Aleutian magmatic arc (Stewart, 1978), and upper Cenozoic sands and sandstones from the forearc basin of the Japan arc (Packer and Ingersoll, 1986). The upper petrofacies is also similar to other Triassic to Lower Jurassic clastic sequences in the Klamath Mountains of central Oregon (Murray and Condie, 1973; Dickinson et al., 1979), and the Methow basin of Washington (Tennyson and Cole, 1978).

Minor changes do occur between the upper green clastics member of the Tyaughton Formation and the Last Creek Formation: (1) the per cent of definite plutonic fragments increases from 1-3% in the upper green clastics member to 1-7% in the Last Creek Formation (Fig. 51K, L, M); (2) microplutonic fragments, likely hypabyssal volcanic detritus, increases in the same way as the plutonic fragments; (3) a few metamorphic grains are present in the Last Creek Formation, whereas none is in the upper green clastics member; (4) undulose quartz, which is very subordinate to nonundulose quartz in the upper green clastics member, is subequal to it in a few samples of the Last Creek; and (5) one sample (UT85-185) from the Last Creek Formation is an arkose (Fig. 51N), which shows a clear plutonic signature and plots in the dissected arc field far from the other samples. This sample is from a distinctive, white, coarse grained, sandstone in the Upper

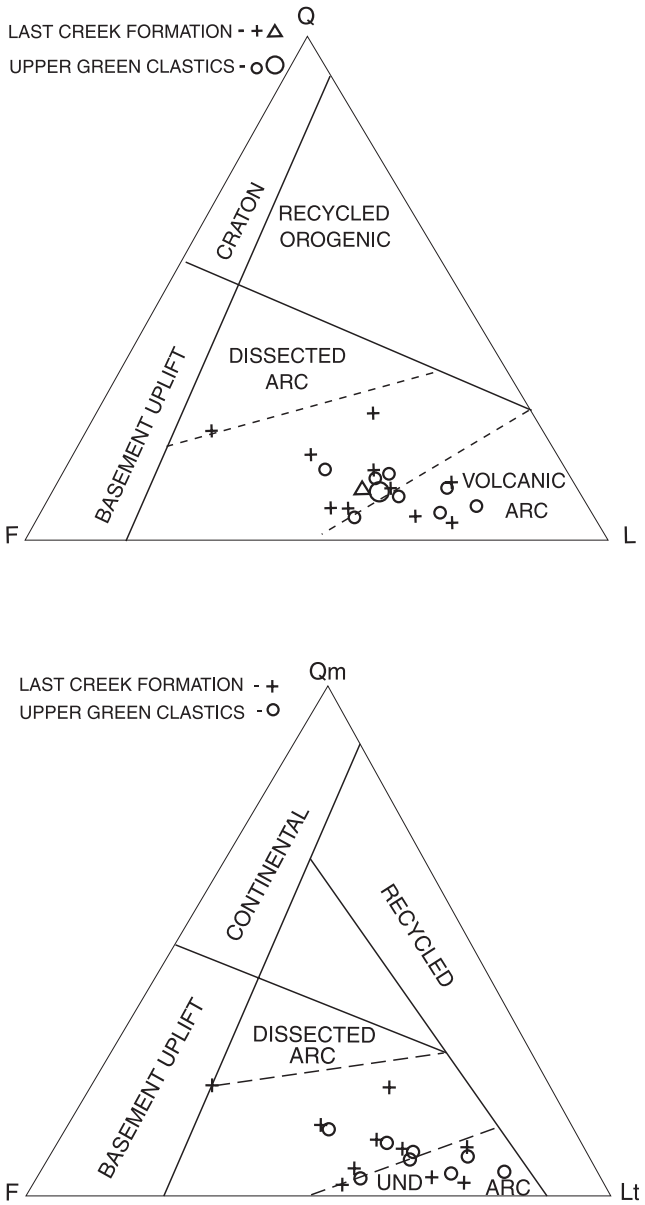


Figure 53. *Provenance plots of upper petrofacies of the Tyaughton and Last Creek formations. Provenance fields after Dickinson et al. (1983). Parameters as in Figure 50. Data in Table 1. UND ARC is the undissected arc field. Large circle and triangle in top plot are mean of each data set.*

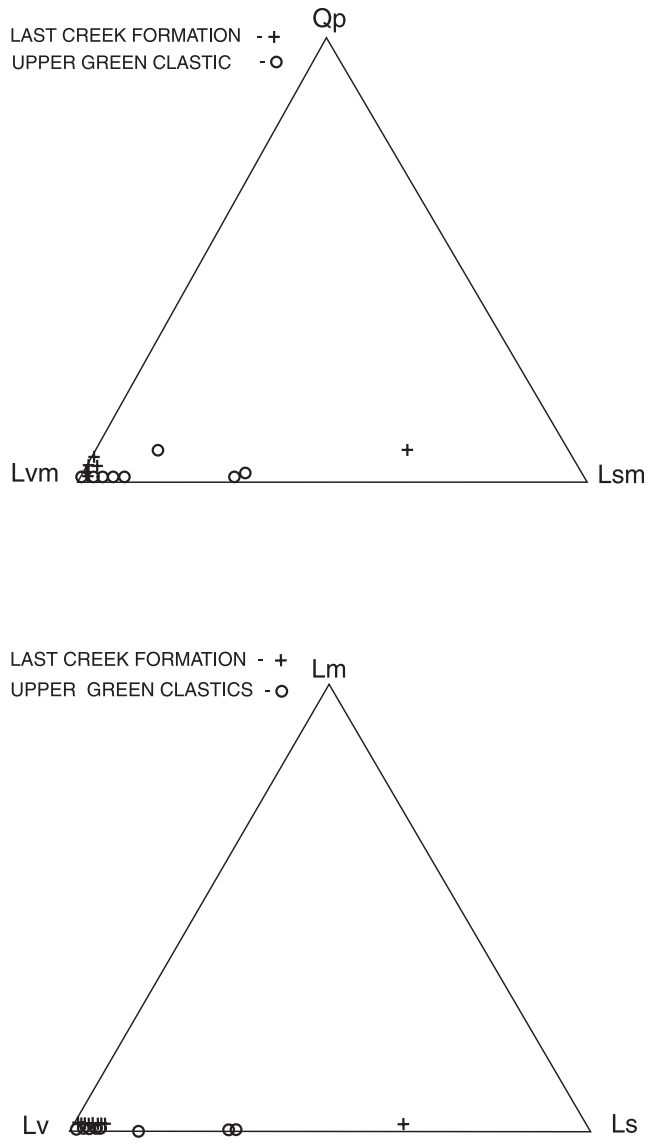


Figure 54. *Provenance plots of lithic clasts of upper petrofacies, Tyaughton Formation and Last Creek Formation. Parameters as in Figure 50. Data in Table 1.*

Hettangian-Lower Sinemurian section. These few arkosic sandstones and the minor changes moving upsection in the upper petrofacies are evidence of the reintroduction of a plutonic and a minor metamorphic source to the basin in Early Jurassic time.

Conglomerate clast lithologies

The lithological character of the conglomerates in all four clastic units generally supports the conclusions drawn from the sandstone petrography. Only a few adequate localities were found for quantitative pebble counts (Fig. 55 and 56), but these were supplemented by numerous localities where approximate counts were made which are probably accurate to within 10% based on the similarity to the quantitiative counts. Two aspects of the conglomerate data from the basal red conglomerate are important for provenance interpretation. First, the pebble-sized limestone clasts are more abundant than in the sandstones (though there are some sandstones in the basal red conglomerate member with limestone clasts, but they were not used for point counting). Conglomerates in some outcrops have up to 95% limestone clasts with minor volcanic clasts. In other places the ratio is reversed and volcanic clasts dominate. Limestone clasts are pink to grey

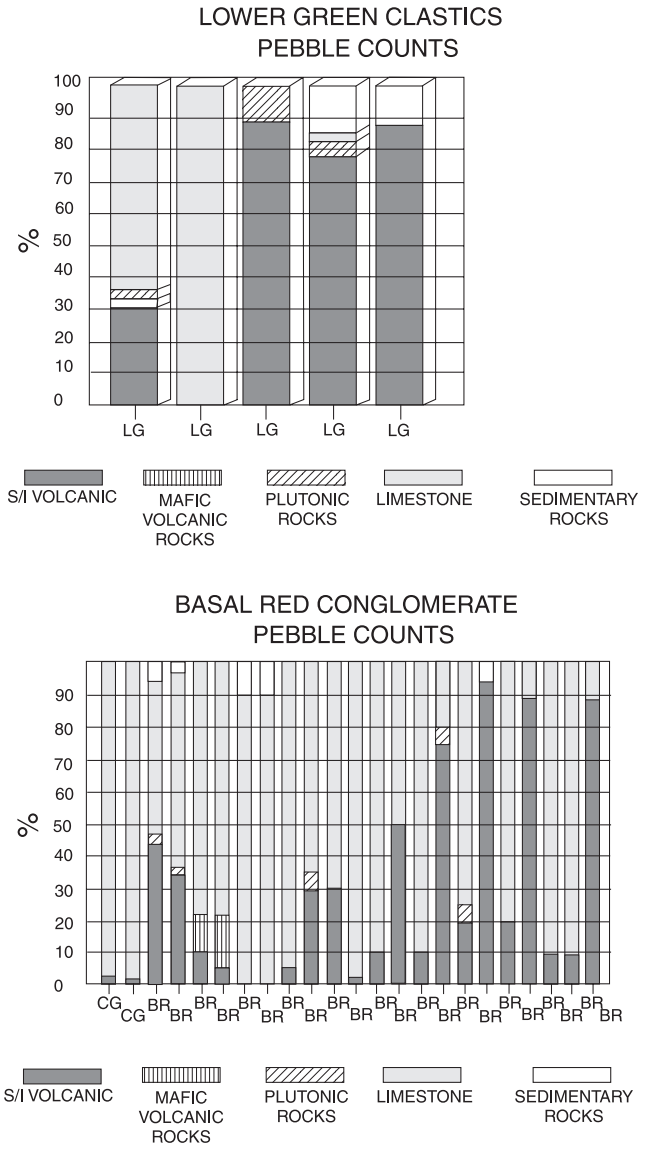


Figure 55. *Conglomerate pebble counts from lower petrofacies, Tyaughton Formation. LG = lower green clastics member; CG = limestone cobble conglomerate member; BR = basal red conglomerate member. S/l volc. = silicic and intermediate volcanic rocks; MAF. VOLC. = mafic volcanics; LS = limestone; SED. RX. = sedimentary rocks.*

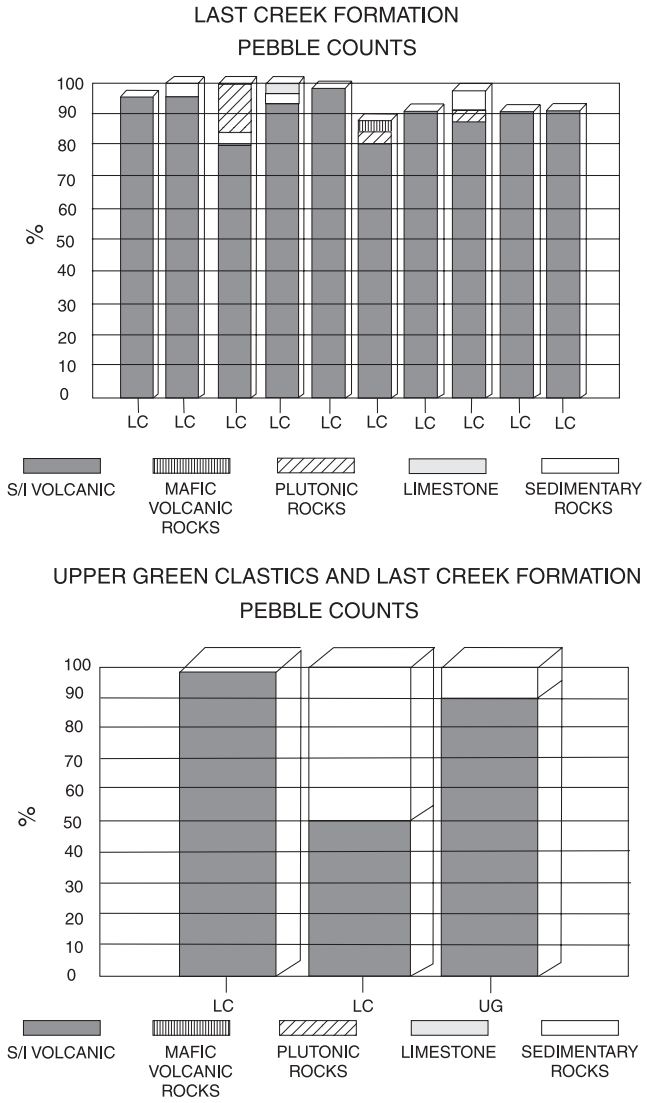


Figure 56. *Conglomerate pebble counts from upper petrofacies, Tyaughton and Last Creek formations. LC = Last Creek Formation, Castle Pass member. UG = upper green clastics member. S/l volc. = silicic and intermediate volcanic rocks; MAF. VOLC. = mafic volcanics; LS = limestone; SED. RX. = sedimentary rocks.*

bioclastic limestones. At three localities conodonts were recovered from clasts which are late Early Norian in age (M.J. Orchard, GSC internal report, OF-1992-3). This age is the same as ages from micrite turbidites in the upper part of the underlying Hurley Formation (and both yield the same species in one case). The coeval conodont ages from clasts in the lower Tyaughton Formation and turbidites in the Hurley Formation supports the interpretation that the basal red conglomerate of the Tyaughton Formation overlies the Hurley Formation, because both units appear to have detritus from the same Lower to Middle Norian shallow-marine limestones; the clasts in the basal red conglomerate member are in a nonmarine unit thus requiring uplift or sea-level fall before deposition. Another possibility is that the Tyaughton limestone clasts are reworked from uplifted Hurley Formation conglomerates that are locally coarse-pebble to cobble sized.

Another noteworthy aspect of the data from the basal red conglomerate member is the difference between the proportions of volcanic and plutonic clasts in the conglomerates and sandstones. Sandstones in the basal red conglomerate member are dominated by plutonic-derived sands. Conglomerates are dominated by volcanic clasts with a minor plutonic component (Fig. 55). This clast-population difference is interpreted as due to the greater susceptibility of intermediate volcanic rocks to mechanical weathering compared to plutonic clasts; therefore, a greater proportion of volcanic clasts is expected in the conglomerates than the sandstones, because the conglomerates were deposited in a more proximal position than the sandstones. In addition, the plutonic rocks predictably would break down into dominantly sand-sized grains because they begin as interlocking, sand-sized, components. In contrast, volcanic rocks originate with only a minor sand-sized component of phenocrysts. Therefore, plutonic rocks will contribute a greater portion of sand to a basin than volcanics per unit volume of rock in the source terrain. This bias toward plutonic clasts would be most expected in arid areas where mechanical weathering is a particularly important process.

From the foregoing reasoning, the conglomerates are a more reliable indicator of the relative lithological character of the source terrain; therefore, the source terrain for the basal red conglomerate member is interpreted to be mainly volcanic, with an important, but minor, plutonic component; the metamorphic component is completely masked in conglomerates.

Conglomerates in the lower green clastics member have the same variable mix of limestone, volcanic rocks, and minor plutonic rocks. Limestone clasts are particularly common as clasts in the basal lower green clastics member where the limestone cobble conglomerate member is missing and then green, limestone-pebble sandstones rest immediately above the massive limestone. Above the lower stratigraphic levels of the lower green clastics member, conglomerates are volcanic-rich with a persistent, minor plutonic component.

Conglomerates in the upper petrofacies are also volcanic-rich (Fig. 56). Intermediate and felsic volcanics are most common with a minor plutonic population. The volcanic-rich nature of the conglomerates mirrors the sandstone data and further suggests that the Tyaughton Formation was deposited on the fringe of an inactive arc.

Conclusions

The clastic rocks of the Tyaughton Formation and Last Creek Formation record the erosion of an inactive magmatic arc. No true volcanic rocks or tuffs were found in the Tyaughton Formation and only a few thin, probable ash beds are in the Lower to Middle Jurassic shales of the Last Creek Formation. The lack of primary volcanic rocks further supports the stratigraphic argument that the Tyaughton Formation was deposited on the Cadwallader Group because Rusmore (1985, 1987) showed that volcanism was active during most of the time the Cadwallader rocks were formed and died out during deposition of the uppermost Cadwallader rocks.

The first phase of deposition of the Tyaughton Formation (Middle to Late Norian) was dominated by plutonic and volcanic detritus. The plutonic component was greater than the volcanic in sandstones (Fig. 57), but the reverse is true in the conglomerates. Taken together, the conglomerate and sandstone data suggest that an inactive volcanic arc in a transitional phase of erosion was the source region. The source area was probably still dominantly volcanic, but with a significant area of plutonic rocks, and minor metamorphic rocks. Shallow-water limestones also were exposed in the source region.

The age of clasts in the basal red conglomerate member demonstrates that some of the bedrock in the source area was only a few million years old when it was eroded and deposited. This evidence, and the proximal depositional setting of the basal red conglomerate member, strongly suggest that the volcanic arc which eroded to form the Tyaughton Formation clastic rocks was the equivalent of the basal Cadwallader Group arc-related basalts (Pioneer Formation), and was the same volcanic arc which supplied detritus to the upper, sedimentary Cadwallader Group. Clasts from the Hurley Formation conglomerates could be the source for some of the lower Tyaughton Group clasts, but many of the more intermediate to felsic plutonic/volcanic source rocks, metamorphic source rocks, and Lower Norian limestone source rocks are not exposed in the Tyaughton Creek region. These “missing” source rocks for the Hurley and Tyaughton formations, if they are to be found at all, are likely to be in another terrane that was originally a part of the same arc as the lower Cadwallader volcanic rocks.

The second phase of deposition, during latest Triassic to Early Jurassic time (late Late Norian to Sinemurian), was almost entirely from erosion of an undissected to mildly dissected volcanic arc (Fig. 57). A minor plutonic component was present in the source area and became a significant part of the basin fill in the Early Jurassic.

One question remains: why did the sudden and dramatic change in petrofacies from a mixed volcanic/plutonic source area to a dominantly volcanic source (Fig. 57) take place at the transition from the *Cassianella* beds member to upper green clastics member? The change may be due to renewed

volcanism in the Cadwallader arc but the lack of primary volcanic rocks in the Tyaughton Formation discounts this option. The stratigraphic position of this transition records a relative drop in sea level, due to either a eustatic fall in sea level or tectonic uplift of the Cadwallader arc and fringing sedimentary rocks. A worldwide review of Upper Norian strata (Hallam, 1981) suggests that the rock record is insufficient to distinguish between these two options. Uplift of the arc would seem to imply even deeper dissection and more of a plutonic signature in the sandstones, not the return to a completely volcanic source, and we have argued that the Tyaughton Formation strata indicate long-term subsidence of the inactive arc, not tectonic uplift.

One possible explanation for the lower to upper petrofacies change is that tectonic uplift of part of the inactive arc changed the drainage pattern such that new, less-dissected regions of the arc were subject to erosion into the fluvial system that fed the sandstones of the upper green clastics member. Taylor et al. (1987) and Taylor and Tajima (1987) have shown that Quaternary uplift in the Vanuatu and Solomon islands arcs in the southwest Pacific has dramatically changed drainage patterns on some volcanic islands over relatively short periods of time.

Nemaia Formation

Data on the provenance of the Nemaia Formation is from a collection of 29 sandstone samples exclusively from the Upper Member of the formation. No samples were collected from the fine grained Lower Member, but the probable tuffaceous nature of the minor sandstones in the Sinemurian rocks supports a volcanic source area.

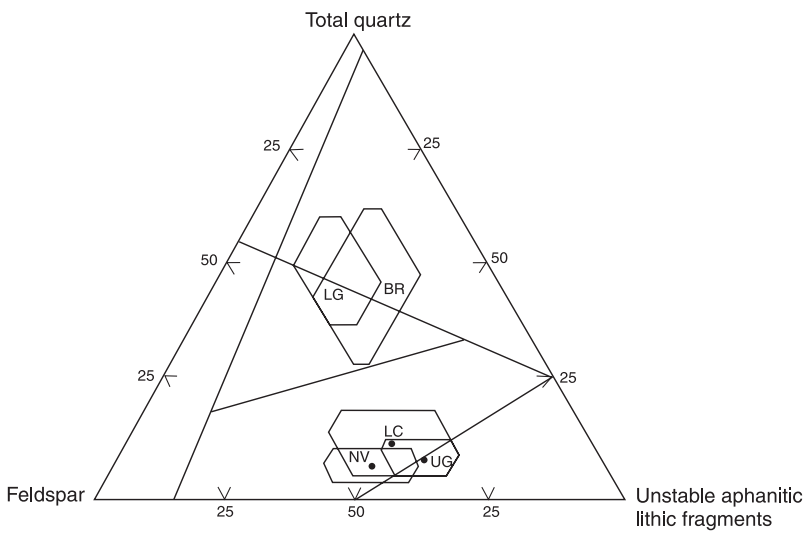


Figure 57.

Summary provenance plots showing mean and one standard deviation area for point counts of sandstone from each unit. Background lines are provenance fields as in Figure 49. BR = basal red conglomerate member, LG = lower green clastics member, UG = upper green clastics member, LC = Last Creek Formation, NV = Nemaia Formation.

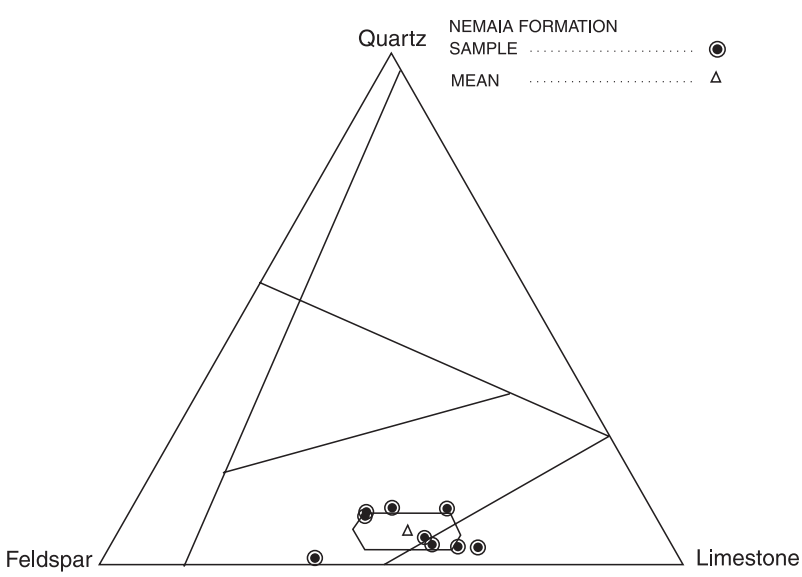


Figure 58.

Summary provenance plot for Nemaia Formation. Triangle is the mean and polygon the first standard deviation of the data set (Q = 6, F = 44, L = 50).

Thin sections from all 29 samples were examined and 9 sandstone thin sections were point counted. Two of the samples are volcanic, a lithic felsite and andesitic tuff, and one sample is an altered tuffaceous sandstone. The methods for point counting were the same as described for the Tyaughton and Last Creek formations, except that the Nemaia Formation thin sections were not stained. Typically 5-15% of the framework grains are altered volcanic rock fragments - the volcanic nature of these grains can be determined but not the type of volcanic rock. Because of this, detailed petrographic analysis is limited.

All sandstones from the upper Nemaia Formation have similar character and comprise a single petrofacies (Fig. 58). The Nemaia petrofacies is similar to the upper petrofacies of the Tyaughton-Last Creek sequence. Both plot in the undissected to transitional arc fields on QFL and QmFLt diagrams (Fig. 57). The sandstones average a QFL of $Q = 6 \pm 3.5$, $F = 44 \pm 9$, $L = 50 \pm 9$. Lithic rock fragments are virtually all volcanic and these range from felsite to lathwork basalts (Fig. 51O, P). Nearly all of the feldspars are plagioclase (Fig. 51O, P), although this is a tentative conclusion because sections were not stained, and up to 40% of the feldspars in any section were too altered to distinguish between plagioclase and K-feldspar. Most of the quartz grains are straight extinction and of volcanic origin. A minor portion of quartz grains are higher undulosity and may be of plutonic origin. A few grains in some sections are polycrystalline quartz of uncertain origin. Between 1-2% of the grains in most sections are quartz or feldspar which are part of a polycrystalline plutonic rock fragment. Conglomerate beds in the Upper Member contain nearly all volcanic clasts, consistent with the sandstones.

In summary, the Nemaia petrofacies probably was derived from a largely undissected volcanic arc with a minor plutonic component. The thin sections confirm the field observation that a few thin beds are tuffs and lithic felsites punctuate the section. We conclude from this data that the Middle Jurassic part of the Nemaia Formation was deposited in a basin adjacent to an active volcanic arc.

FACIES RELATIONS OF THE ROCK UNITS

Implications of the vertical sequence of facies

The changes in depositional environment through time for the Tyaughton Formation and Last Creek Formation are plotted as changes in paleobathymetry in Figure 59 and as a total subsidence curve in Figure 60. Below preliminary conclusions are drawn from these data concerning the relative importance of eustatic sea level and tectonic movements for the history of the stratigraphic sequence in an attempt to gain insight into the tectonic setting.

Biostratigraphic resolution in this sequence is approximately equal to the duration of a single ammonoid zone, which is on the order of one million years for the Upper Triassic and Lower Jurassic (Van Hinte, 1976). This value of resolution would be accurate if the section did not include disconformities, but of course the Tyaughton Formation has two distinct depositional interruptions (Fig. 7) so that any calculated sediment accumulation rate is a minimum value. Major disconformities are not well documented within the Last Creek Formation, but structural complications render the thickness estimates crude for the upper part of the unit. The sediment accumulation rates in Figure 59 assume that the Tyaughton Formation members span most of the late Late Norian and that the disconformities represent relatively short intervals of time.

The entire sequence does not represent a simple response to eustatic sea level changes (Fig. 59) as presented by Haq et al. (1987), although two aspects of the Tyaughton-Last Creek section are compatible with the sea level curve. The erosional disconformity at the Triassic-Jurassic boundary, which separates the two units, corresponds to the lowest position of sea level in the entire Mesozoic and Cenozoic before the Holocene lowstand (Haq et al., 1987), and the transgressive nature of the Last Creek Formation is mirrored by a gradual rise of sea level of a similar duration (Fig. 59).

The magnitude of postulated sea level rise in the Early Jurassic (about 100 m, Haq et al., 1987) is about the amount represented by a change from nearshore to outer shelf environment, as in the Last Creek Formation. The subsidence curve in Figure 60 suggests that the total subsidence of the upper part of the Cadwallader terrane was greater than the postulated rise of sea level by an order of magnitude. The overall shape of the subsidence curve is similar to thermal subsidence beginning in the Early Norian.

The apparently short-duration disconformities in the Tyaughton Formation are more problematic. The change from alluvial redbeds to marine limestones (Fig. 59) requires either a transgression or tectonic subsidence. The sudden change from *Monotis*-bearing, offshore, limestones to fluvial clastic rocks, derived from the same limestone unit, suggests either a rapid regression or tectonic uplift. In each case, neither explanation can be ruled out because the minimum relative vertical change required to produce the disconformities is only tens of metres which either sea level change or tectonic movements can accommodate in tens of thousands of years. The main trend within the disconformity-bounded sequences in the Tyaughton Formation is from nonmarine to marine or toward deeper-water facies (Fig. 59). One explanation for the patterns in the Tyaughton Formation is that short-term sea level changes are disrupting a general response to long-term thermal subsidence due to the cooling of the Cadwallader volcanic arc, which had been inactive since the Early Norian (Rusmore, 1987).

It is suggested from the previous observations that the Tyaughton Formation and Last Creek Formation record both the effects of short-term sea level changes in the Late Triassic and the long-term fall of sea level until the end of the Triassic and rise of sea level in the Early to Middle Jurassic, all superimposed on a long-term thermal subsidence of the basin. Thermal subsidence throughout this period is compatible with the combined physical stratigraphic evidence of the Cadwallader terrane (Cadwallader Group, Tyaughton Formation, and Last Creek Formation). The Cadwallader Group records the termination of arc-related volcanism in the

Early Norian (Rusmore, 1987) and there is no evidence of contemporary volcanism throughout the time of deposition of Tyaughton Formation and Last Creek Formation, except for possible thin ash beds in Jurassic strata (Rusmore, 1987; Rusmore et al., 1988; see "Biostratigraphy" section). This evidence implies that the Cadwallader volcanic arc, in the narrow sense of those rocks presently exposed in the southeastern part of this study area remained inactive from Early Norian to early Middle Jurassic, or about 40 million years. If the subsiding arc followed typical thermal subsidence rates for oceanic to transitional crust (50 to 35 m/Ma after the initial few million years of subsidence, Sclater et al., 1971), then about 200-300 m of thermal subsidence would have occurred during deposition of Tyaughton Formation and approximately 1000 m of subsidence during deposition of Last Creek Formation. These values are compatible with the estimate for subsidence of the two units (Fig. 60). A sea level fall of 50-100 m during Tyaughton Formation deposition (Haq et al., 1987; Fig. 59) would partly counteract the effect of thermal subsidence on the Tyaughton Formation and implies that a net, long-term, relative sea level change of approximately 100 to 200 m occurred during deposition of the Tyaughton Formation. The sea level rise of 100 m in the Lower Jurassic would add to the approximate 1000 m of postulated thermal subsidence during deposition of the Last Creek Formation.

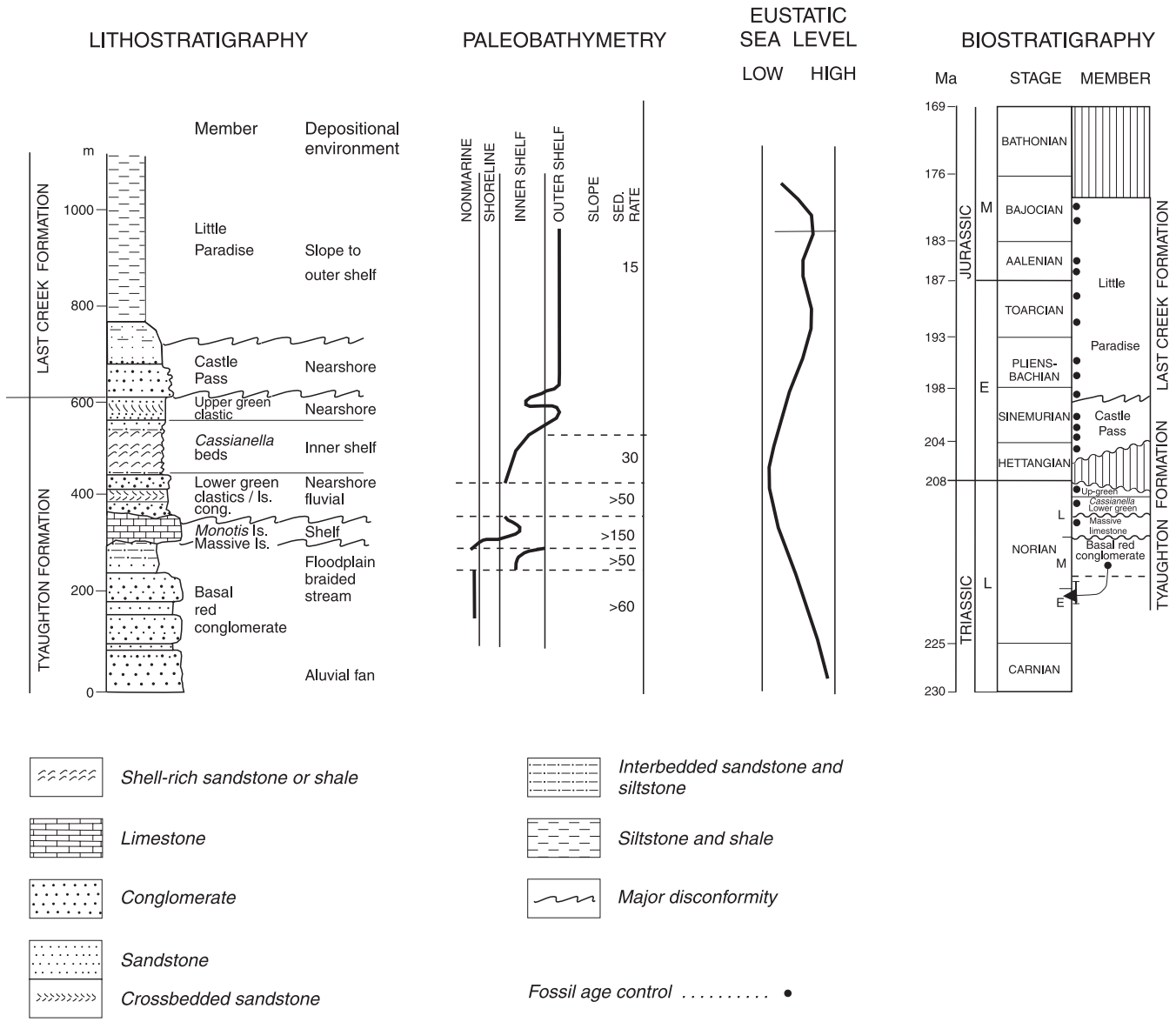


Figure 59. *Paleobathymetry of Tyaughton and Last Creek formations from interpretations of depositional environment in this bulletin. Eustatic sea level curve after Haq et al. (1987). Limitations of sedimentation rates are discussed in text. Vertical ruled pattern represents hiatuses; Ls, ls = limestone; cong. = conglomerate. CAR. = Carnian; NOR. = Norian; HETT. = Hettangian; SIN. = Sinemurian; PLIEN. = Pliensbachian; TOAR. = Toarcian; AAL.= Aalenian; BAJ. = Bajocian; BATH. = Bathonian;*

Implications of the lateral relations of facies

The lateral relationship of facies between the Castle Peak area and the Chilko Lake area is summarized here, but for all, except the lower Middle Jurassic part of the sequence, the Chilko rocks are too poorly exposed and sparsely fossiliferous to reach well constrained conclusions. The relation of the Triassic to Jurassic rocks discussed in this report to coeval rocks outside the Chilcotin Ranges is discussed in a later section (see "Comparison to adjacent basinal terranes").

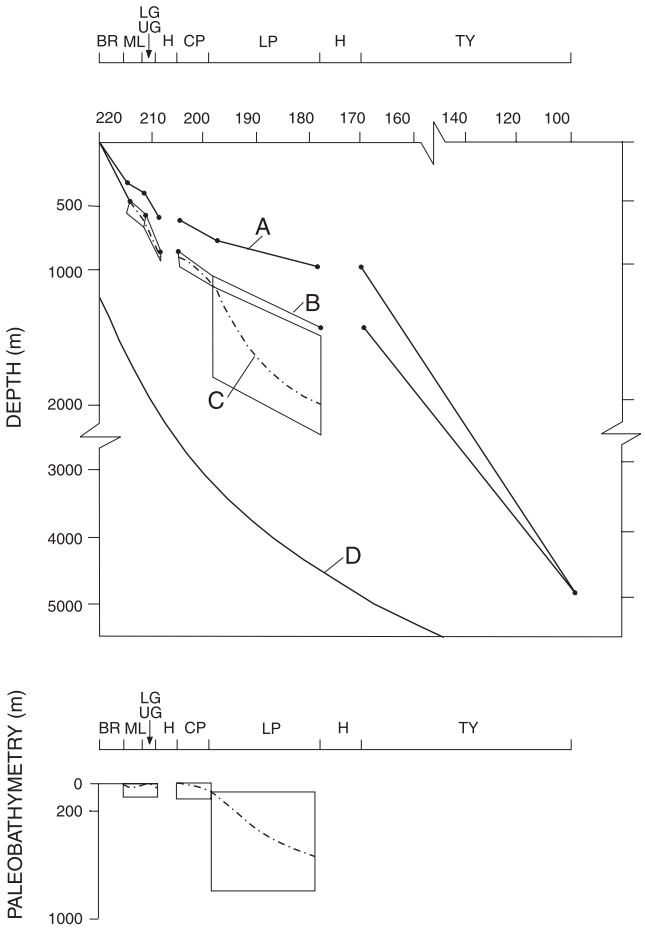


Figure 60. *Subsidence curve for the Tyaughton and Last Creek formations (units 1 to 6). Curve A is the simple sediment accumulation through time as measured by present-day thickness of units, B is the curve corrected for compaction, and C is corrected for paleobathymetry using the curve in Figure 59 and the evidence presented in the text. The method used was after Van Hinte (1976). The subsidence curve (C) would produce a tectonic subsidence curve with the same shape, but with a magnitude which was about 30-50% of the subsidence curve shown here. The shape of the subsidence curve most closely resembles a thermal subsidence curve (D). Unit abbreviations are: BR = Basal red conglomerate, ML = Massive limestone, LG/UG = Lower green and upper green clastics, H = hiatus, CP = Castle Pass, LP = Little Paradise, TY = Tyaughton basin units.*

The paucity of outcrops of Triassic rocks in the Chilko Lake area preclude knowledge of how the Tyaughton Formation changes outside the main outcrop belt near Castle Peak. The general lithologies found along Tatlayoko Lake are broadly similar to those found in the main area of the Tyaughton Formation. This suggests that nonmarine to shallow marine units were deposited near an inactive arc in the latest Triassic throughout the Chilcotin Ranges.

The Nemaia Formation disconformably overlies the Tyaughton Formation and Hurley Formation, which is similar to the local erosional relation of the Last Creek Formation over the Tyaughton Formation in the southeastern part of the belt. The lower part of the Nemaia Formation is exposed in discontinuous, fault-bounded sequences. No Hettangian strata have been found but one location has yielded Lower Sinemurian fossils. From this sparse data, it appears that Lower Sinemurian strata become finer grained from southeast to northwest in the Chilcotin Ranges. This may indicate that the subsidence to deeper marine conditions recorded in the Castle Peak area within the Sinemurian section occurred earlier to the west, or that the more northwestern areas represent deeper parts of the Lower Jurassic basin.

Most of the Lower to Middle Jurassic (Upper Sinemurian to Aalenian) Last Creek and Nemaia formations are similar sequences of shale and siltstone with minor sandstone. The Lower Jurassic shale sequences in both units are punctuated by a pulse of Pliensbachian conglomerates, which indicates basin-wide input of gravels, perhaps a relative fall of sea level. There is a modest eustatic fall of sea level in the Pliensbachian on the sea-level curve of Haq et al. (1987). There is more sandstone and conglomerate in the upper Nemaia Formation than in the Last Creek Formation. Middle Jurassic (Bajocian) strata show a significant change from shales with minor sandstone in the eastern Last Creek Formation to a sandstone-rich, coarsening-upward sequence with minor volcanic rocks in the Nemaia Formation to the west that is interpreted to be the record of a prograding delta on the margin of an active volcanic arc. (Too few observations were made of paleocurrent indicators to give a better estimate of the source direction for the delta or shoreline in the Bajocian.) These major changes from the Last Creek Formation to Nemaia Formation suggest that the Last Creek Formation was isolated, or distal, from the Middle Jurassic volcanic arc and the source area for the coarse clastic rocks and volcanic rocks of the upper part of the Nemaia Formation. Further, the location of renewed volcanic activity was not in the study area, but in another terrane. We discuss the possible sources of this Middle Jurassic volcanism in the next section.

TECTONOSTRATIGRAPHIC RELATIONSHIP OF CHILCOTIN RANGES ROCK UNITS TO OTHER TERRANES IN SOUTHWESTERN BRITISH COLUMBIA

Introduction

The rock units of this report can be separated into two tectonostratigraphic packages, the Upper Triassic rocks and Lower to Middle Jurassic rocks, each bounded by known or

probable unconformities. We and other workers have argued that the Upper Triassic ((?)middle to late Norian) Tyaughton Formation was probably deposited on the Upper Triassic (Carnian to middle Norian) Cadwallader Group (Rusmore, 1987; Rusmore et al., 1988; Umhoefer, 1990). In previous papers, the Cadwallader, Tyaughton, and Last Creek rock units (and by correlation, the Nemaia Formation of this report) comprise the Cadwallader terrane. We retain this terrane definition but note that in some of the possible tectonic models presented below the Lower Jurassic units are part of a widespread overlap sequence covering two or more terranes and the Cadwallader terrane is best confined to the Triassic Cadwallader and Tyaughton rock units. In other models, part of what we call the Cadwallader terrane is included in other terranes.

In this concluding section we will briefly compare the rocks of the Cadwallader terrane with the other major terranes of southwestern British Columbia and speculate on their relationships: the Wrangellia terrane on Vancouver Island and the Queen Charlotte Islands, the Stikine terrane of the southern Intermontane belt (mainly north of the Chilcotin Ranges, but also southwest of Chilko Lake); and the Harrison Lake, Bridge River, Methow, and Quesnellia terranes on the east side of the Coast belt and within the Intermontane belt (Fig. 61, 62).

It is premature to present a definitive conclusion on the paleotectonic relation of these terranes because (i) the detailed depositional and tectonic setting and age are not known for many stratigraphic sequences and igneous rock bodies in the region and (ii) the many complex deformational events that occurred after Middle Jurassic time are not sufficiently known to reconstruct the spatial relations between the terranes. Most specifically, the latter point includes the uncertainty in estimates of the amount of slip on major dextral strike-slip faults of Late Cretaceous to early Tertiary age such as the Fraser, Yalakom, and related faults; uncertainty in the amount of shortening in major thrust belts of mid-Cretaceous age within and on the margins of the Coast belt; and the possibility of large-scale sinistral strike-slip offsets that are only approximately constrained in location and amount of offset (Monger et al., 1994). These structural and/or tectonic

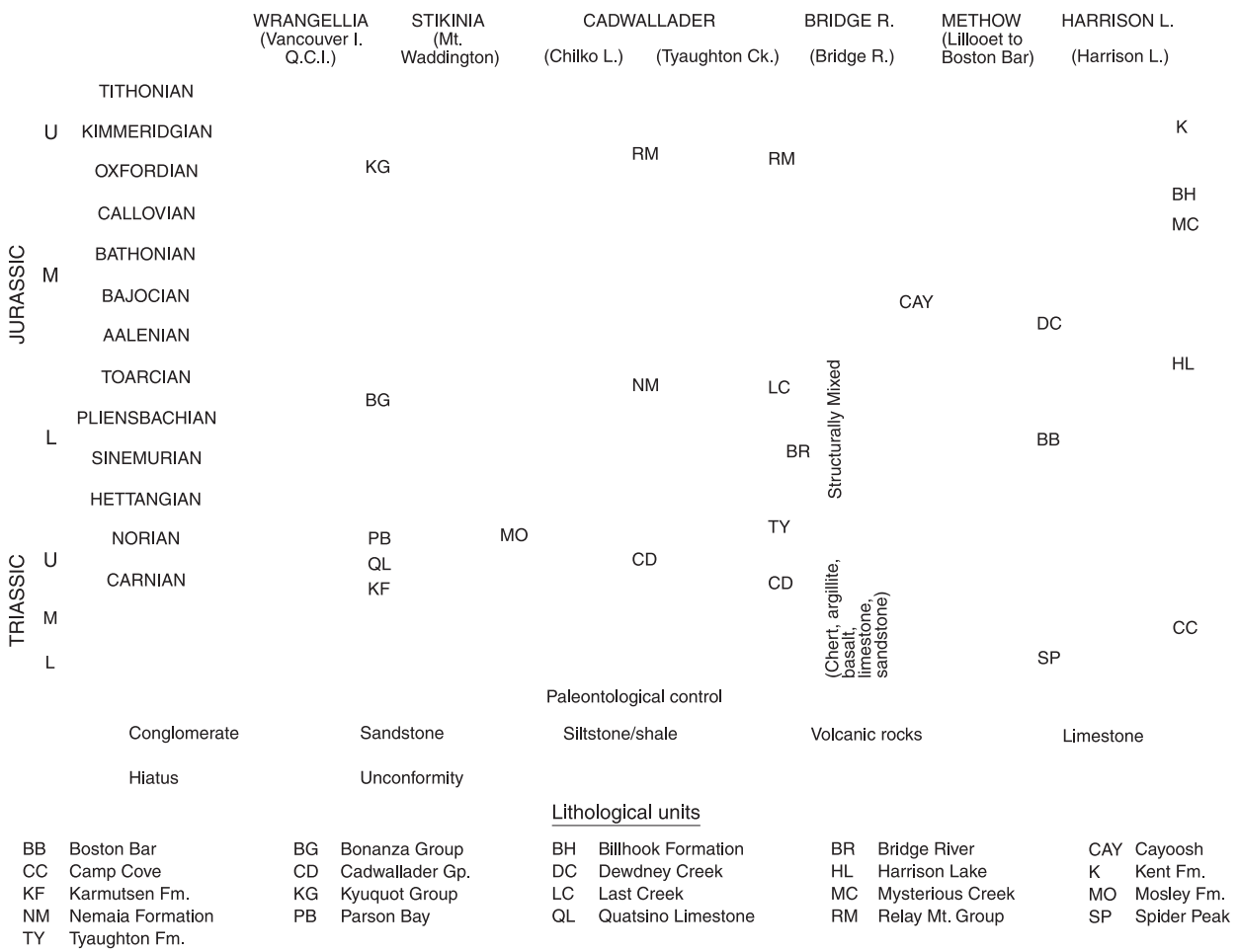


Figure 61. *Schematic stratigraphic sections of principal areas in southwestern British Columbia with Triassic and Jurassic rock units. See text for discussion of sections.*

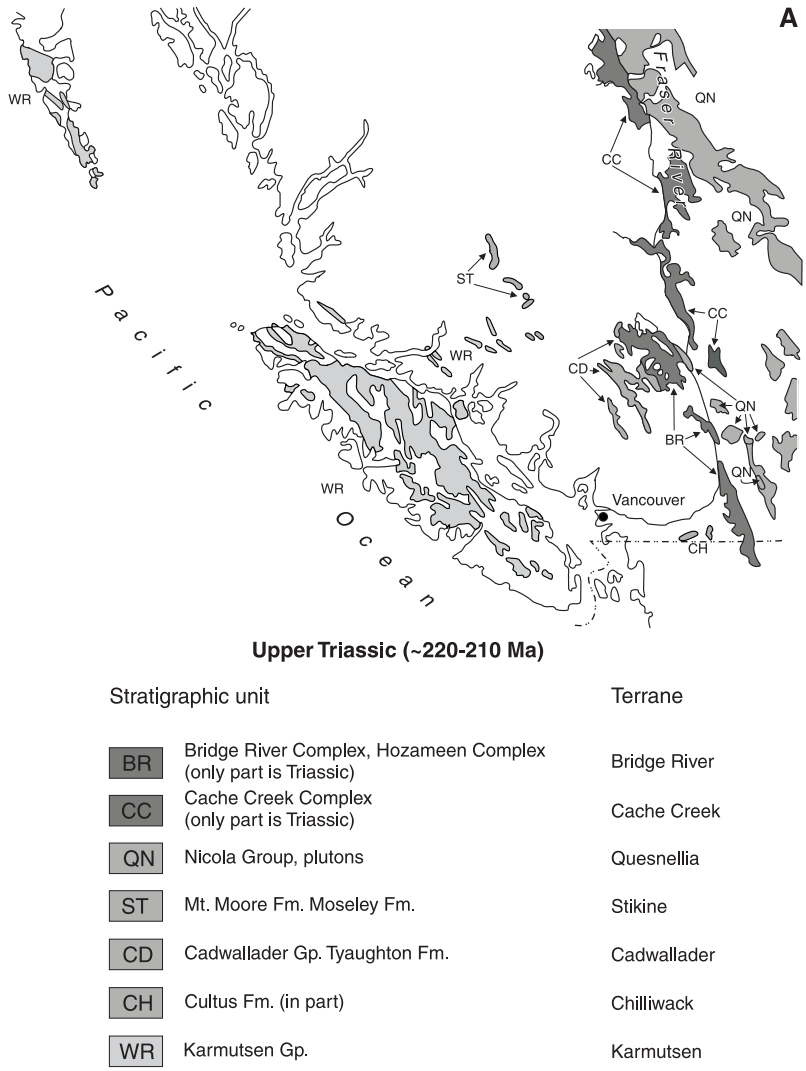


Figure 62. *Three maps of southwestern British Columbia depicting the distribution of (**A**) Upper Triassic, (**B**) Lower Jurassic, and (**C**) lower Middle Jurassic rocks (modified after Wheeler and McFeely, 1991). See text for discussion and Figure 61 for more information on the stratigraphy of individual terranes.*

problems are the focus of much ongoing research, the results of which will have a great impact on the spatial relations of the older Mesozoic terranes.

The general conclusion about fault restorations that can be made at this time is that at least 100 km and most likely 200 to 300 km of dextral slip must be restored on the Yalakom and Fraser fault systems (Kleinspehn, 1985; Monger, 1985; Umhoefer and Schiarizza, 1993); this implies that the Cadwallader terrane and any terranes to the west of it should be restored the same distances to the southeast relative to more eastern terranes. Shortening across the Coast belt is likely to be many hundred kilometres (e.g. McGroder, 1991; Journeay and Friedman, 1993), which would restore the Wrangellia terrane much farther west than it presently lies relative to the Cadwallader and more eastern terranes.

The qualitative conclusion that in Late Triassic to Early Jurassic time the more western terranes in southwestern British Columbia were at least hundreds of kilometres farther west and south then their present position relative to eastern terranes is supported by paleobiogeographical studies. Upper Triassic macrofossils (Tozer, 1982a) and Lower Jurassic (Pliensbachian) ammonites (Tipper, 1984; Smith and Tipper, 1986; Smith et al., 1988) from the Wrangellia, Stikine, and Quesnellia terranes suggest that those terranes were at the approximate latitude of Nevada in the Late Triassic to Early Jurassic.

The detailed summary of the displacement record of the major fault systems in southwestern British Columbia, and therefore the relative spatial distribution of terranes, is beyond the scope of this paper. For this reason and the

Lower Jurassic (Sinemurian)(~200 Ma)

Stratigraphic unit		Terrane
BR	Bridge River Complex, Hozameen Complex (only part is Lower Jurassic)	Bridge River
CC	Cache Creek Complex (in part)	Cache Creek
QN	Quesnel River Gp. (in part), plutons	Quesnellia
CD	Last Creek Fm., Huckleberry Fm. (in part)	Cadwallader
CH	Cultus Fm.	Chilliwack
WR	Bowen Island Group (in part) Bonanza Group (in part) Kunga Group (in part)	Wrangellia

Figure 62. *(cont.)*

fragmentary nature of our knowledge of fault reconstructions of the region, we choose to present only the general relation between terranes and suggest some of the possible tectonic settings that would explain the distribution of tectonic assemblages.

Tectonostratigraphic summary of Cadwallader terrane

The Cadwallader Group and Tyaughton, Last Creek, and Nemaia formations comprise the Cadwallader terrane as we are defining it. These units span Late Triassic to early Middle Jurassic time (Rusmore, 1987; this report) and are mainly found in the Chilcotin Ranges, but a belt of Upper Triassic volcanic rocks near Pemberton has been included in the Cadwallader Group (Fig. 61, 62) (Riddell, 1991; Journeay, 1993). In contrast to our definition, Riddell et al. (1993) assign Nemaia Formation to the Methow Terrane.

The Cadwallader Group has a sequence of basalts at the base interpreted from geochemical data to have formed in a volcanic arc (Rusmore, 1987). The basalts are overlain by late Carnian to early Norian volcaniclastic and deep marine clastic rocks, which record the termination of volcanic activity in the early Norian and deposition of arc-derived turbidites (Rusmore, 1987).

The Tyaughton Formation of this report is interpreted to overlie the Cadwallader Group sedimentary rocks, but everywhere the contact is a fault (Umhoefer, 1990). An unconformity is implied between the two units, because the basal unit of the Tyaughton Formation is alluvial-fluvial conglomerates and sandstones, whereas the inferred underlying Cadwallader rocks are marine turbidites. The Tyaughton Formation is a sequence of clastic rocks and carbonates deposited in nonmarine and shallow marine environments after volcanic activity had ceased. The source area for the clasts was an eroding volcanic arc and the upper level plutonic roots of the arc: we assume that the Cadwallader arc was the source, but no intermediate plutonic rocks like the plutonic-derived clasts are known from the Cadwallader terrane.

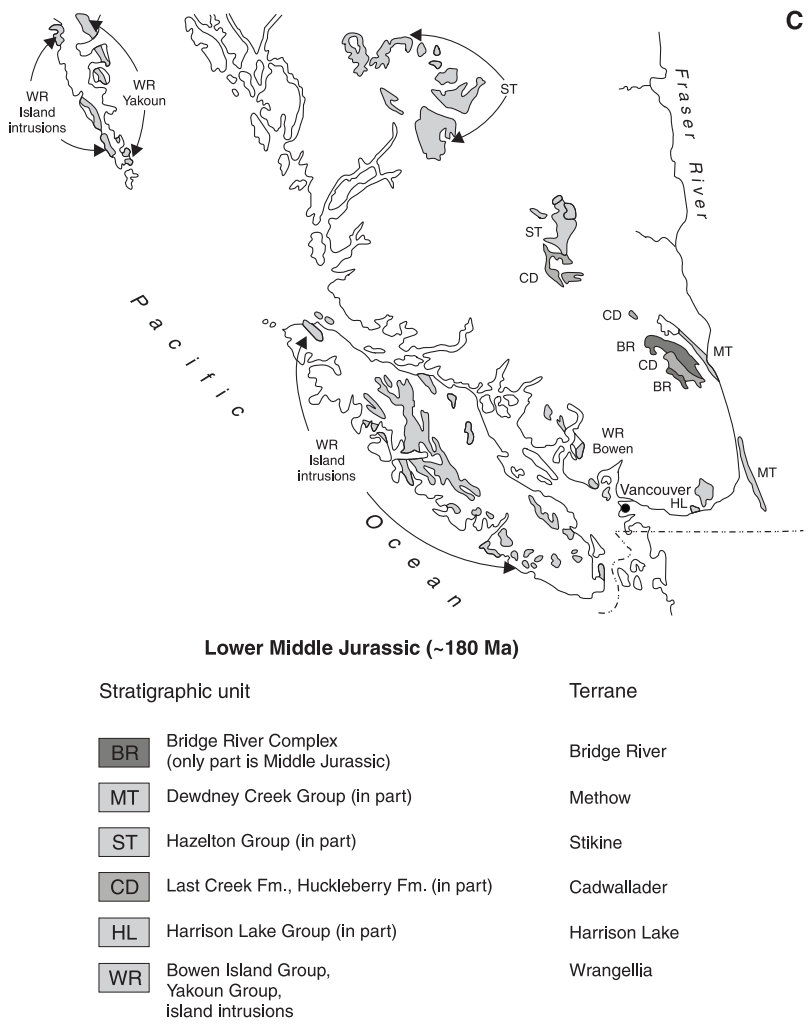


Figure 62. *(cont.)*

The Last Creek Formation was demonstrably deposited on the Tyaughton Formation. The Last Creek Formation records nearshore deposition of conglomerates and sandstones in the Hettangian, followed by a steady fining upward of the clastic rocks through the Sinemurian, which we interpret to be a deepening of the depositional environment to at least middle to outer shelf. The conditions may have been deeper, but the abundant ammonite fauna favors shelf conditions. The Last Creek Formation continues to be a shale-dominated marine section up through the Bajocian rocks and we suggest outer shelf to slope conditions prevailed. An exception to this is the Lower Pliensbachian sequence of conglomerates and sandstones, which abruptly overlies the upper Sinemurian shales and is interpreted to indicate an erosional or nondepositional interval (or both) due to a sea level fall or tectonic-induced uplift. The dominantly volcanic nature of the sandstone grains, with a minor amount of plutonic grains, suggests that the erosion of the extinct (?)Triassic arc that began during the deposition of the Tyaughton Formation continued during deposition of the Last Creek Formation. A few thin probable ashy shale beds indicate this part of the basin was far from concurrent volcanic activity.

The Nemaia Formation appears to be similar to the Last Creek Formation in its poorly studied lower member (Lower Jurassic), with dominantly fine grained marine rocks punctuated by a Lower Pliensbachian coarse grained sequence. The Hettangian and lowest Sinemurian was not found in the Nemaia Formation. The lower Middle Jurassic is a distinct contrast to the Last Creek Formation. In the Aalenian to early Bajocian, the Nemaia Formation was the site of a prodelta to outer delta front or submarine fan progradation seen in a progressive coarsening-up sequence. Marginal marine conditions are never recorded, but instead the basin returns to an upper prodelta setting at the top of the Nemaia Formation. This coarsening upward delta sequence includes minor, but important tuffs. We interpret the Nemaia Formation to be an Early Jurassic offshore to deep marine basin distal to a volcanic arc followed in the Aalenian to early Bajocian by a deltaic setting adjacent to an active volcanic arc.

In summary, the Cadwallader terrane was first the site of a Carnian to early Norian arc. The basin was the marginal marine fringe of an inactive arc in the latest Triassic. It then progressively became a deeper marine setting, with only

minor indication of far away volcanic activity in Early Jurassic time. Finally, the terrane formed the flank of an early Middle Jurassic arc.

Comparison to adjacent basinal terranes

Here we compare the Cadwallader terrane history to the record from other terranes of southwest British Columbia (Fig. 61, 62) and give the most probable conclusions on relations amongst terranes. The comparison is made for three discrete times, late Carnian-early Norian, Sinemurian, and Bajocian (Fig. 62).

The Late Triassic was a time of diverse tectonic assemblages in southwest British Columbia. The Carnian to early Norian arc of the Cadwallader terrane is broadly similar to volcanic arc activity in southern Stikinia in the Mount Waddington map area, but many differences exist in detail (Rusmore and Woodsworth, 1991a; Umhoefer et al., 1994). Both terranes have dominantly Carnian to early Norian volcanism, but volcanic activity dies completely in the Cadwallader terrane whereas tuffs in the upper Norian of the Stikinia sequence of the Mount Waddington area attest to continued volcanism (Rusmore and Woodsworth, 1991a; Mustard and van der Heyden, 1994). It is permissible that these two terranes were part of the same volcanic arc, but their different latest Norian history demands that they were at least separate parts of an arc. It is also possible that the two terranes were parts of different arcs. Quesnellia terrane has a distinct Late Triassic volcanic arc that was west facing (Mortimer, 1987) and continued into the latest Triassic and earliest Jurassic, in contrast to the Cadwallader and Stikine arcs. This relationship indicates that the Cadwallader (with or without Stikine arc) was further offshore relative to North America than the Quesnellia arc.

In contrast to these arc terranes, the Bridge River terrane was a deep oceanic basin and an accretionary complex in the Triassic (Schiarizza et al., 1990; Archibald et al., 1991). Wrangellia in the Triassic was an oceanic basaltic plateau and overlying deep marine basin as represented by the thick Karmutsen basalts and overlying argillites (e.g. Jones et al., 1977). The Harrison Lake and Methow terranes have no Late Triassic record. Besides the west-facing Quesnellia volcanic arc, which may have been the northern oceanic equivalent of the continental Triassic arc of the southwest United States (Mortimer, 1987), the spatial relation of the other terranes is difficult to assess. It is known from fossil evidence that the Wrangellia, Stikinia, and Cadwallader terranes were deposited in the northeastern Pacific basin in the Triassic, south of where they are now (Tozer, 1982a).

The Early Jurassic record in southwest British Columbia starts with only two areas of Hettangian rocks, in the Queen Charlotte Islands (Tipper and Guex, 1994) and the Last Creek Formation of the Chilcotin Ranges. These occurrences are both nearshore deposits and may represent two sides of the same basin, but the scanty Hettangian record in the region precludes a firm interpretation.

The Sinemurian is a distinct contrast to the Upper Triassic, because for the first time the available data allows for a simple interpretation of the entire region. This interpretation is that the Bonanza magmatic arc of Wrangellia on Vancouver Island (Armstrong, 1988) has a large backarc basin represented in part by the deep marine, fine grained clastic sequences of the Cadwallader terrane. The minor ashy shales and tuffs in the Last Creek and Nemaia formations support this model. This interpretation implies that the diverse and therefore probably far separated Upper Triassic terranes of southwest British Columbia were in close enough proximity by the Sinemurian to form a coherent arc-backarc system. Eastern Quesnellia was also an arc at this time, but the relation between it and the possible Wrangellia-Cadwallader arc-backarc is uncertain. We can infer that if the Sinemurian Quesnellia arc was west facing like the Triassic arc below it, then it was a completely different arc system than the more western Wrangellia-Cadwallader arc. The Cache Creek ocean-accretionary complex may have lain between the two arcs in this model (Monger et al., 1982).

The possible occurrence of the accretionary oceanic Bridge River terrane between the Cadwallader and Wrangellia terranes (Cordey and Schiarizza, 1993), argues against the above arc-backarc model in the Sinemurian. The alternative is that the Wrangellia volcanic arc lay on a separate plate to the west of the Bridge River and Cadwallader terranes. The lack of strata of this age in Harrison Lake and Methow terranes (Mahoney, 1994) makes it difficult to evaluate the model rigorously.

The Bajocian is much better represented in the terranes of southwest British Columbia than the older strata (Fig. 61, 62). Three models for the configuration of the terranes are viable. It is clear that in the present configuration, there is an early Middle Jurassic magmatic belt along the southern Coast Belt that is probably part of a volcanic arc and includes parts of the Harrison Lake and Wrangellia terranes (Mahoney, 1994; Friedman and Armstrong, 1995). There is also a belt of intermediate volcanic rocks and deep marine sedimentary rocks (Dewdney Creek Group) deposited adjacent to a west-facing arc in the Methow terrane (Mahoney, 1992). The two belts are interpreted to be either two separate volcanic arcs or a single arc that was cut and displaced by Early Cretaceous sinistral faulting (Monger et al., 1994).

The two-arc model has two variations in which the position of the Cadwallader terrane is the main variable. Mahoney (1994) and one of the present authors (Tipper) point out that the Bajocian Nemaia Formation changes from coarse clastic rocks with minor tuffs to a fine grained clastic section in Last Creek Formation. They conclude that this trend represents a west to east transition from more proximal to distal arc-derived sediments that formed on the east side of the Aalenian to Bajocian Wrangellia-Harrison Lake volcanic arc (Mahoney, 1994). In this model, the Cayoosh assemblage and Bridge River Complex represent the middle of the backarc basin and lay east of the Cadwallader terrane, and the Methow terrane was farther to the east adjacent to an undetermined volcanic arc (Journeay and Mahoney, 1994).

The alternative interpretation of the early Middle Jurassic Cadwallader terrane is implied by the correlation across the Yalakom Fault after 115 km of dextral offset (Riddell et al., 1993). Riddell et al. (1993) included the Nemaia Formation in the Methow terrane and correlated it to the Middle Jurassic Dewdney Creek Group of the northern Methow terrane in the Camelsfoot Range. They noted the correlation of a series of faults on both sides of the Yalakom fault that include the same rock units in fault slices. The fault restoration from these correlations produces an approximate northeast to southwest change from the more arc-proximal Dewdney Creek Group to the less proximal Nemaia Formation to the distal Last Creek Formation, which presumably lay southwest of the same arc. This interpretation also implies that the Bridge River terrane was west of the Cadwallader terrane in the Middle Jurassic and was the accretionary complex between the eastern and western arcs.

Monger et al. (1994) suggested that there was just one arc, not two in southwest British Columbia in early Middle Jurassic. They suggested that the two belts of Middle Jurassic magmatic rocks were originally one and the current western belt was displaced in the late Early Cretaceous southward by a large-scale sinistral fault system that lies between the Cadwallader-Bridge River and Harrison Lake terranes. This model implies that the Cadwallader terrane was part of the forearc of the west-facing Methow arc, and far south of the Wrangellia-Harrison Lake terranes, which were a forearc-arc pair of the same arc.

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