# Nina Creek Group and Lay Range Assemblage, north-central British Columbia: remnants of late Paleozoic oceanic and arc terranes[1]

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**Abstract**: In north-central British Columbia, a belt of upper Paleozoic volcanic and sedimentary rocks lies between Mesozoic arc rocks of Quesnellia and Ancestral North America. These rocks belong to two distinct terranes: the Nina Creek Group of the Slide Mountain terrane and the Lay Range Assemblage of the Quesnel terrane. The Nina Creek Group is composed of Mississippian to Late Permian argillite, chert, and mid-ocean-ridge tholeiitic basalt, formed in an ocean-floor setting. The sedimentary and volcanic rocks, the Mount Howell and Pillow Ridge successions, respectively, form discrete, generally coeval sequences interpreted as facies equivalents that have been interleaved by thrusting. The entire assemblage has been faulted against the Cassiar terrane of the North American miogeocline. West of the Nina Creek Group is the Lay Range Assemblage, correlated with the Harper Ranch subterrane of Quesnellia. It includes a lower division of Mississippian to Early Pennsylvanian sedimentary and volcanic rocks, some with continental affinity, and an upper division of Permian island-arc, basaltic tuffs and lavas containing detrital quartz and zircons of Proterozoic age. Tuffaceous horizons in the Nina Creek Group imply stratigraphic links to a volcanic-arc terrane, which is inferred to be the Lay Range Assemblage. Similarly, gritty horizons in the lower part of the Nina Creek Group suggest links to the paleocontinental margin to the east. It is assumed that the Lay Range Assemblage accumulated on a piece of continental crust that rifted away from ancestral North America in the Late Devonian to Early Mississippian by the westward migration of a west-facing arc. The back-arc extension produced the Slide Mountain marginal basin in which the Nina Creek Group was deposited. Arc volcanism in the Lay Range Assemblage and other members of the Harper Ranch subterrane was episodic rather than continuous, as was ocean-floor volcanism in the marginal basin. The basin probably grew to a width of hundreds rather than thousands of kilometres.

**Résumé** : Dans le centre nord de la Colombie-Britannique, une ceinture de roches volcaniques et sédimentaires du Paléozoïque supérieur repose entre les roches mésozoïques de l'arc de la Quesnellie et le protocontinent nord-américain. Ces roches appartiennent à deux terranes distincts : le Groupe de Nina Creek du terrane de Slide Mountain et l'Assemblage de Lay Range du terrane de Quesnel. Le Groupe de Nina Creek est composé de roches s'échelonnant du Mississippien au Permien tardif, incluant des argilites, cherts et basaltes tholéiitiques de dorsale médio-océanique déposés sur un plancher océanique. Les roches volcaniques et sédimentaires, c'est-à-dire les successions de Mount Howell et de Pillow Ridge, respectivement, forment des séquences généralement contemporaines mais distinctes, interprétées comme étant des faciès équivalents qui furent intercalés par chevauchement. Tout l'assemblage a été fragmenté par le jeu des failles contre le terrane de Cassiar du miogéoclinal de l'Amérique du Nord. À l'ouest du Groupe de Nina Creek apparaît l'Assemblage de Lay Range, il est corrélé avec le sous-terrane de Harper Ranch de la Quesnellie. Il inclut une division inférieure constituée de roches sédimentaires et volcaniques qui s'échelonnent du Mississippien au Pennsylvanien précoce, certaines affichant une affinité continentale, et une division supérieure qui représente un arc insulaire datant du Permien, avec des tufs et laves basaltiques contenant des grains détritiques de quartz et de zircon originaires du Protérozoïque. Les horizons tufacés dans le Groupe de Nina Creek impliquent l'existence de liens stratigraphiques avec un terrane d'arc volcanique, présumé être l'Assemblage de Lay Range. De même, les horizons gréseux dans la portion inférieure du Groupe de Nina Creek suggèrent la présence de liens avec la marge paléocontinentale à l'est. On suppose que l'Assemblage de Lay Range fut édifié sur un morceau de croûte continentale, qui dérivait en s'éloignant du protocontinent nord-américain aux temps du Dévonien tardif à Mississippien précoce, par la migration en direction ouest d'un arc face à l'ouest. C'est l'extension d'un arrière-arc qui a produit le bassin marginal de Slide Mountain, dans lequel le Groupe de Nina Creek fut déposé. Le volcanisme d'arc dans l'Assemblage de Lay Range et pour les autres membres du sous-terrane de Harper Ranch était intermittent plutôt que continu, il en était de même pour le volcanisme sur le plancher océanique dans le bassin marginal. L'élargissement du bassin fut plus probablement de l'ordre de centaines plutôt que de milliers de kilomètres.
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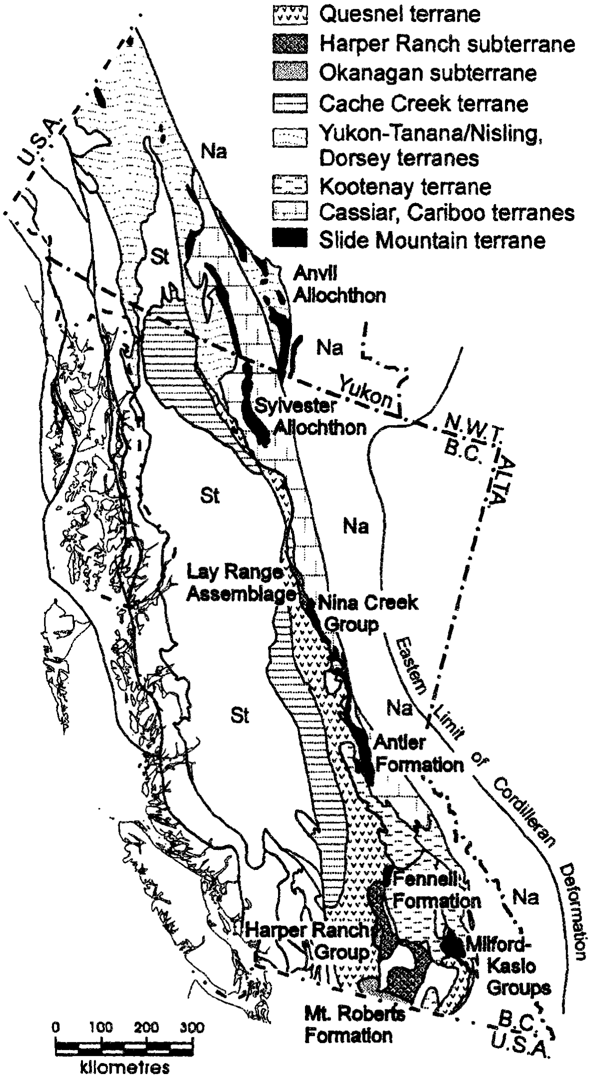
## Introduction

In north-central British Columbia, two different assemblages of upper Paleozoic volcanic and sedimentary rocks separate Mesozoic arc rocks of Quesnellia from the miogeocline of Ancestral North America. The easternmost assemblage comprises the Mississippian to Permian Nina Creek Group, composed predominantly of basalt, gabbro, fine-grained siliciclastics, chert, and mafic to ultramafic bodies. It represents the Slide Mountain terrane, which can be traced intermittently along the length of the Canadian Cordillera and nearly everywhere is in thrust contact with pericratonic or displaced miogeoclinal rocks (Fig. 1). Parts of the Slide Mountain terrane have been interpreted as remnants of a large late Paleozoic ocean (Monger 1977; Harms 1986; Gabrielse 1991) or as a marginal basin along the late Paleozoic continental margin (Monger 1977; Klepacki and Wheeler 1985; Schiarizza and Preto 1987; Struik 1987; Miller et al. 1984, 1992; Nelson 1993; Roback et al. 1994).

The other upper Paleozoic terrane is the Mississippian to Permian Lay Range Assemblage, a package of mixed sedimentary and volcanic rocks overlain by a thick succession of arc-derived basaltic volcanics. This assemblage has been assigned to the Harper Ranch subterrane, the type area of which is in the southern Cordillera, where it comprises mafic to felsic volcanics and fine to coarse siliciclastics and limestones (Fig. 1). Harper Ranch rocks are treated as a subterrane of Quesnellia, because they are stratigraphically overlain by early Mesozoic arc volcanics and sediments characteristic of this terrane (Monger et al. 1991). The Harper Ranch subterrane and its probable equivalents have been interpreted either as a fringing arc along the western margin of Ancestral North America (Tempelman-Kluit 1979; Miller 1987; Gabrielse 1991; Nelson 1993; Roback et al. 1994) or as a far-traveled arc system (Speed 1979; Snyder and Brueckner 1985; Brueckner and Snyder 1983). It is everywhere structurally outboard (to the west of) or above the Slide Mountain terrane. This relationship also occurs in the southern part of the Lay Range Assemblage. To the north, however, it is faulted directly against the Cassiar terrane of the North American miogeocline, with no intervening Nina Creek Group preserved.

This paper presents a summary of geologic data gathered during 5 years of mapping in the Omineca Mountains of north-central British Columbia along a previously poorly understood portion of the boundary between Ancestral North America and accreted rocks of the eastern Intermontane superterrane. The area lies between the relatively recently described segments of this boundary in the Cassiar Mountains (Nelson 1993; Harms 1986) and the type areas of the Slide Mountain terrane in southern British Columbia (Struik 1987; Schiarizza and Preto 1987). This paper describes the stratigraphy of the Nina Creek Group and Lay Range Assemblage, the geochemistry of their igneous rocks, and the structural and stratigraphic relationships between them and adjacent terranes. Given their proximity and generally coeval stratigraphy, the formation of the Slide Mountain terrane and Harper Ranch subterrane may be related. The occurrence of tuffaceous deposits within pelagic sediments of the Nina Creek Group constitutes an indirect, but important link between these two packages. Data presented in this paper will be integrated with regional information to suggest that the Slide Mountain terrane represents an upper Paleozoic back-arc basin that formed behind the adjacent Harper Ranch west-facing arc.

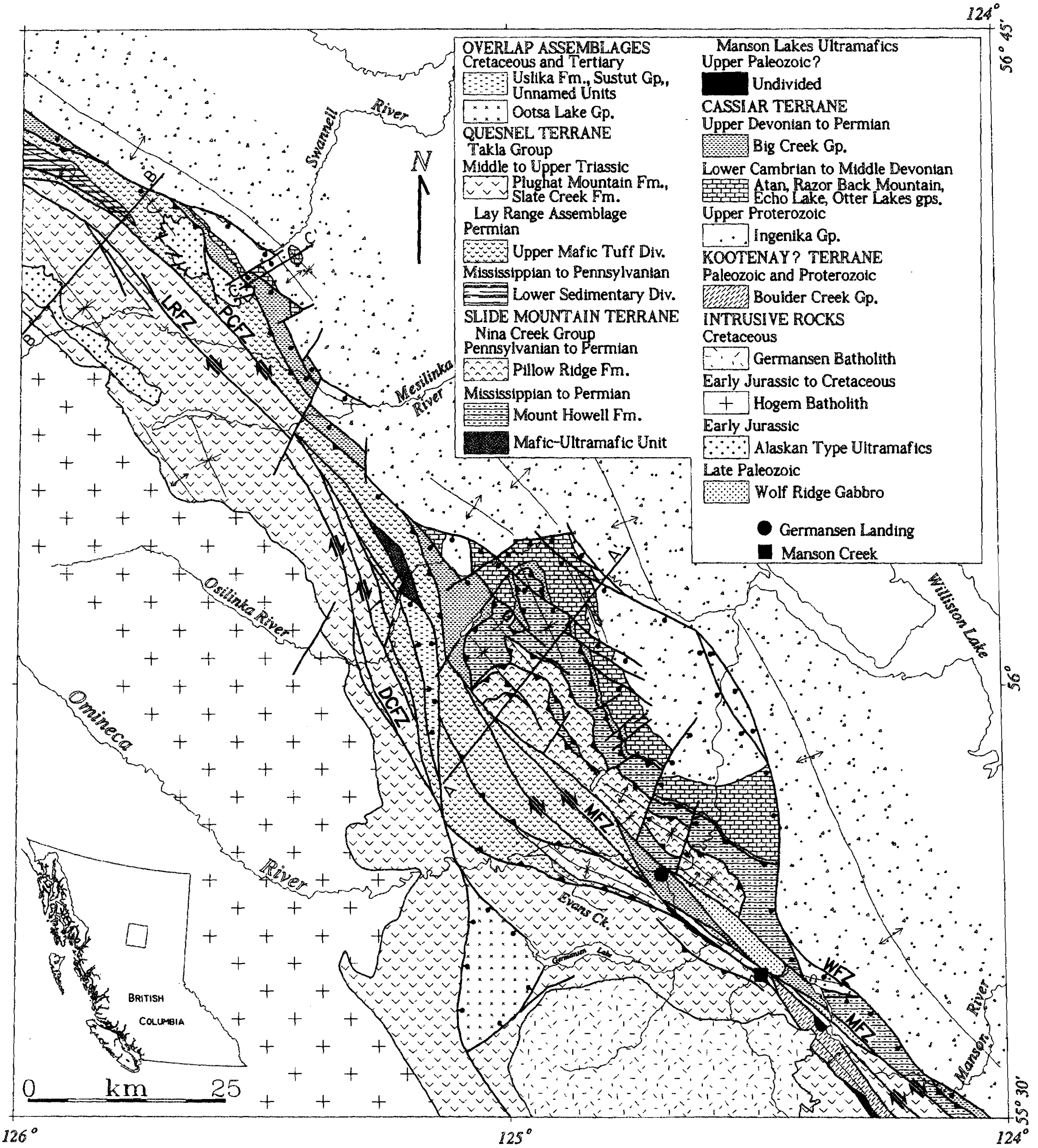
**Fig. 1.** Simplified tectonic assemblage map of the Canadian Cordillera outlining principal terranes. Na, Ancestral North America; St, Stikine terrane.



## Geological setting

Rocks of the Nina Creek Group and the Lay Range Assemblage extend from the Manson River to the northern end of the Lay Range, a strike length of over 200 km (Fig. 2). Nina Creek rocks were originally mapped as part of the Cache Creek Group by Armstrong (1949), but were subsequently shown to be more typical of oceanic rocks found along the eastern part of the Intermontane Belt, now known as the Slide Mountain terrane (Monger and Patterson 1974; Monger 1977; Gabrielse 1975; Monger et al. 1991). Upper Paleozoic rocks in the Lay Range were first recognized by Roots (1954) and were later placed within the Eastern Assemblage (Monger 1977), a term used before the Slide Mountain terrane was coined. These rocks were named the Lay Range Assemblage by Ross and Monger (1978) and were tentatively

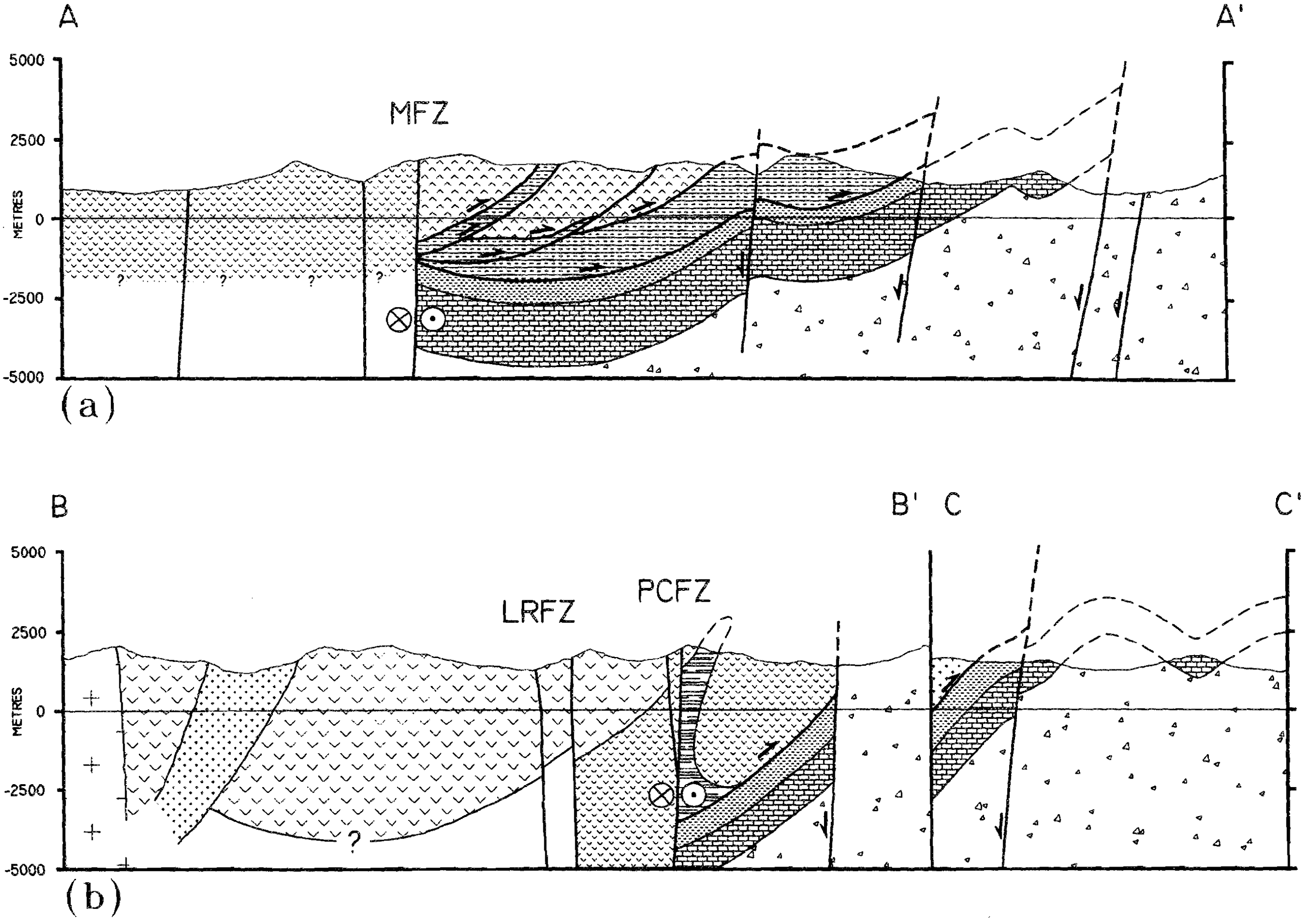
**Fig. 2.** Simplified geologic map of the area between the Swannell and Manson rivers. Locations of cross sections in Fig. 3 are shown by lines A–A′, B–B′, and C–C′. MFZ, Manson fault zone; DCFZ, Discovery Creek fault zone; LRFZ, Lay Range fault zone; PCFZ, Polaris Creek fault zone; WFZ, Wolverine fault zone.



reassigned to the Late Devonian to Late Permian Harper Ranch subterrane of Quesnellia by Monger et al. (1991).

Parts of the Nina Creek Group and Lay Range Assemblage have been thrust onto miogeoclinal and rift-related siliciclastics and carbonates of Late Proterozoic to Early Permian age (Figs. 3–5) (Ferri and Melville 1994; Ferri et al. 1992, 1993; Gabrielse 1975). The latter rocks belong to the Cassiar and Kootenay terranes, which represent displaced parts of the Ancestral North American continental margin (Ferri and Melville 1994). West of the upper Paleo-

**Fig. 3.** Schematic cross sections within the map area. (*a*) Section A–A′, across the northern trace of the Nina Creek Group. (*b*) Sections B–B′ and C–C′, across the Lay Range. See Fig. 2 for locations and unit designations. ⊗, fault block moving away from observer; ⊙, fault block moving towards observer.



zoic rocks are lower Mesozoic arc-related volcanic and sedimentary rocks of Quesnellia, represented in this region by the Upper Triassic to Lower Jurassic Takla Group. They are intruded by the Jura-Cretaceous Hogem Batholith and the Cretaceous Germansen Batholith. Both the Takla Group and the Lay Range Assemblage are intruded by Triassic to Jurassic Alaskan-type ultramafic bodies, some of which are believed to be the plutonic equivalents of the Takla volcanics (Irvine 1976; Nixon et al. 1990).

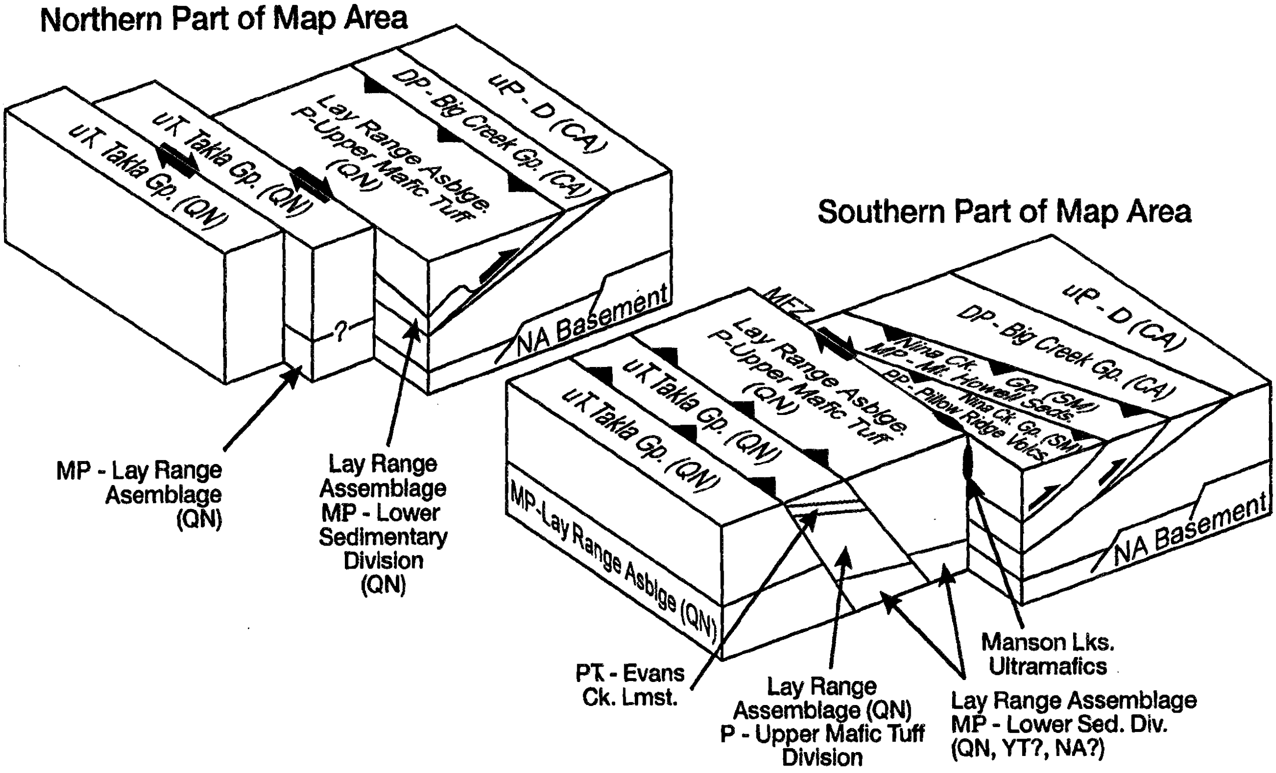
The Nina Creek Group consists of a stack of relatively thin thrust sheets that is separated by a near-bedding-parallel thrust fault from the underlying Cassiar terrane (Fig. 3*a*). The top of the footwall in this region consists of the Devonian to Permian Big Creek Group, composed of black, fine-grained siliciclastics and minor felsic tuff, correlated, in part, with the rift-related Earn Assemblage (Gordey 1991). The Nina Creek thrust panels, and the underlying Cassiar strata, exhibit little internal deformation, but have been broadly folded into upright to southwest-verging structures. The actual imbrication within the Nina Creek sequence is probably more complicated than shown (Figs. 2, 3*a*). However, the monotonous lithologies and the paucity of fossil data do not allow further delineation of thrust repetitions.

In contrast, the structural style of the Lay Range Assemblage is characterized by tight folding and faulting (Fig. 3*b*). Bedding generally dips steeply to the southwest and many sections are overturned; one major, tight, northeasterly overturned fold has been mapped southwest of the Swannell River. In this area, the assemblage has been thrust on top of Cassiar rocks that exhibit moderate large-scale upright to southwest-verging folds that postdate emplacement. Polyphase deformation in Cassiar rocks was accompanied by metamorphism, which reached amphibolite grade at deeper structural levels. Metamorphism within Nina Creek and Lay Range rocks is subgreenschist or lower greenschist grade.

The Manson, Discovery Creek, Lay Range, and Polaris Creek fault zones (Fig. 2) are en echelon, northwest-trending, strike-slip fault systems of Late Cretaceous to early Tertiary age. These fault systems are concurrent and subparallel with other regional-scale right-lateral fault structures, such as the McLeod Lake, Pinchi, Finlay, and northern Rocky Mountain Trench fault systems (Gabrielse 1985).

Slivers of serpentinite, talc-serpentine schist, and listwaenite are found within the Manson and Polaris Creek fault zones or along flat-lying, northeasterly verging thrusts above continental rocks tentatively assigned to the Kootenay terrane (Ferri and Melville 1994). Although no direct link with the

**Fig. 4.** Block diagram illustrating structural and stratigraphic relationships in the map area. CA, Cassiar terrane; NA, Ancestral North America; YT, Yukon–Tanana terrane; SM, Slide Mountain terrane; QN, Quesnel terrane.



Nina Creek Group can be demonstrated, these ultramafic bodies are included with the Slide Mountain terrane, owing to their inferred oceanic or upper mantle origin. The late Paleozoic or early Mesozoic Wolf Ridge Gabbro also occurs in the Manson fault zone. It may represent pieces of upper oceanic lithosphere and, as such, be part of the Slide Mountain terrane. A more thorough description of the geology of the region is given by Ferri and Melville (1994).

## Nina Creek Group

The Mississippian to Permian Nina Creek Group is subdivided into two informal units: the Mount Howell succession, which is typified by deep water siliceous sediments, gabbro, and basalt, and the Pillow Ridge succession, which is predominantly basalt (Fig. 6). Strata of both successions generally dip to the southwest and have been structurally imbricated, forming composite sections up to 7000 m thick. Conodonts extracted from cherts of the Mount Howell succession range in age from Mississippian to Late Permian. Fossil data, although limited, from beds within Pillow Ridge basalts indicate an Early Pennsylvanian to Permian age range. The two successions are thus largely coeval and are inferred to be facies equivalents that have been imbricated by thrusting. Nowhere have they been seen in stratigraphic contact, although gabbro sills near the top of the Mount Howell succession are compositionally similar (see section on Geochemistry of igneous rocks) to basalts of the Pillow Ridge succession, suggesting that they are part of a feeder system to overlying volcanics that were later removed by thrusting and replaced by another section of Pillow Ridge basalts.

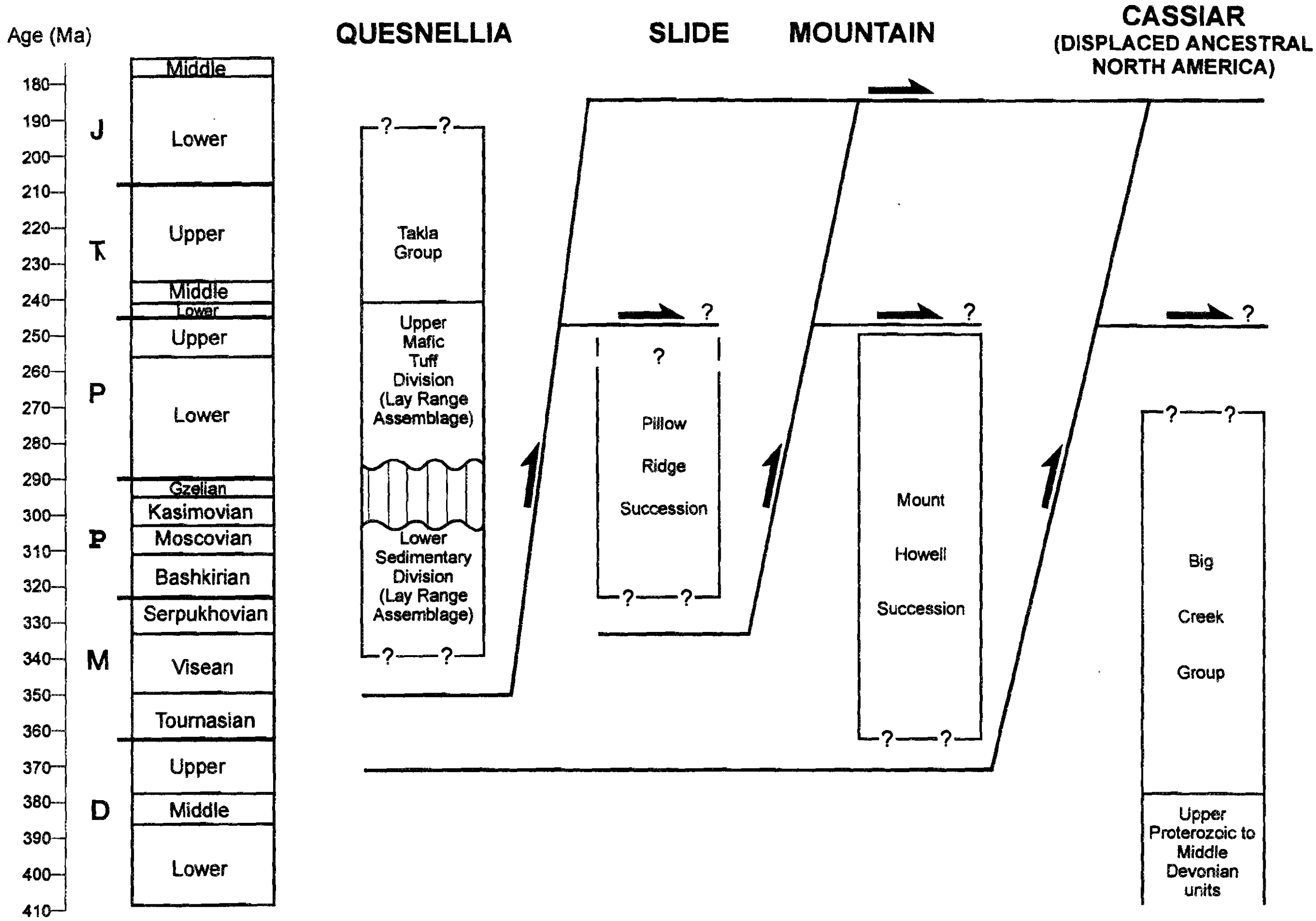
### Mount Howell succession

Mount Howell lithologies are typified by thin to moderately bedded argillites, siliceous argillites, varicoloured cherts, and ribbon cherts. Maroon- and salmon-coloured cherts occur only in the uppermost part of the section. Gabbroic sills up to 1000 m thick intrude the upper parts of the Mount Howell succession at various stratigraphic levels, and some are traceable for several kilometres. They are finely to coarsely crystalline, with subequal pyroxene and strongly altered (sericitized) plagioclase. Minor constituents of the succession are basaltic flows, tuffaceous sediments, sandstone, thin and discontinuous micritic limestone layers, and a quartz-bearing tuff or porphyry towards the base of the unit. Basalt sections up to 50 m thick become more important south of the Omineca River in the upper part of the Mount Howell succession.

Tuffaceous sediments, polymictic sandstones and coarse-grained quartz sandstones, wackes, and conglomerates are a minor component of the Mount Howell succession, but they are important because they suggest the proximity of continental as well as volcanic sources. Tuffaceous sediments are thickly bedded and occur in sections up to 5 m thick, bounded gradationally by typical argillites of the Mount Howell succession. Coarser fractions contain crystal fragments of feldspar and pyroxene (Fig. 7). The tuffaceous sediments appear to become volumetrically important towards the south, where they are interbedded with coarse-grained quartz sandstone.

Locally, polymictic volcanic sandstone, wacke, and conglomerate are interlayered with argillites and tuffaceous sediments. They are composed of subrounded to rounded clasts

Fig. 5. Time–space history of the various terranes within the map area.



of chert, quartz, carbonate, argillite, relatively unaltered feldspar–pyroxene-porphyritic volcanic rock, metamorphic rocks (as shown by undulose and sutured polycrystalline quartz grains), and rare, broken crystals of potassic and calcic feldspar (Fig. 8). Their composition suggests a mixed continental – immature volcanic provenance.

Some thin-bedded isolated sandstone or wacke subunits, however, are neither interbedded with nor contain volcanic material. They are found immediately south of the Osilinka River and comprise well-rounded quartz and chert grains, and appear to have been derived from a mature continental terrane (Fig. 9).

Cherts within the Mount Howell succession have yielded conodonts of Mississippian, Pennsylvanian, and Permian ages. The localities are scattered and allow only a broad age range estimate. Mississippian conodonts were recovered from the base of the sequence south of the Osilinka River, whereas Early to Middle Pennsylvanian and Permian conodonts were found elsewhere. Permian fossils were generally recovered from the upper, gabbro-bearing sections. A 1000 m thick thrust sheet within Pillow Ridge basalts north of the Omineca River contains Early to Middle Pennsylvanian conodonts at its base and Guadalupian (Late Permian) conodonts in its upper part. Although fossil data suggest a general younging upwards through various sections of the Mount Howell succession, there is not enough fossil control along the belt to preclude repetitions by cryptic thrust faults, as has been demonstrated in similar rocks of the Antler Formation in the northern Cariboo Mountains (Struik and Orchard 1985).

### Pillow Ridge succession

The Pillow Ridge succession is at least 5000 m thick northwest of Nina Lake, and is dominated by basalt with lesser sediments, tuff, and gabbro sills. Over 95% of the Pillow Ridge succession is composed of variolitic, pillowed, or massive basalt, which contains irregular bodies of fine-grained gabbro. The monotonous nature of this unit, coupled with the lack of fossil control, does not preclude the presence of cryptic thrust faults, which have structurally thickened the unit. Massive and interpillow basaltic breccia is also common. Very fine grained to fine-grained phenocrysts of feldspar, pyroxene, and olivine are found locally in the basalts. The groundmass is composed of titanaugite, plagioclase, and opaque minerals. Low-grade metamorphic minerals include chlorite, epidote, clinozoisite, prehnite, and pumpellyite, all of which occur as groundmass or vein and vesicle fillings.

Argillite and chert intervals from 1 to 10 m thick occur within the basalts and are similar to those in the Mount Howell succession. Cherts may also be thickly bedded to massive, and in many localities they form irregular masses up to 10 m in diameter within the basalt.

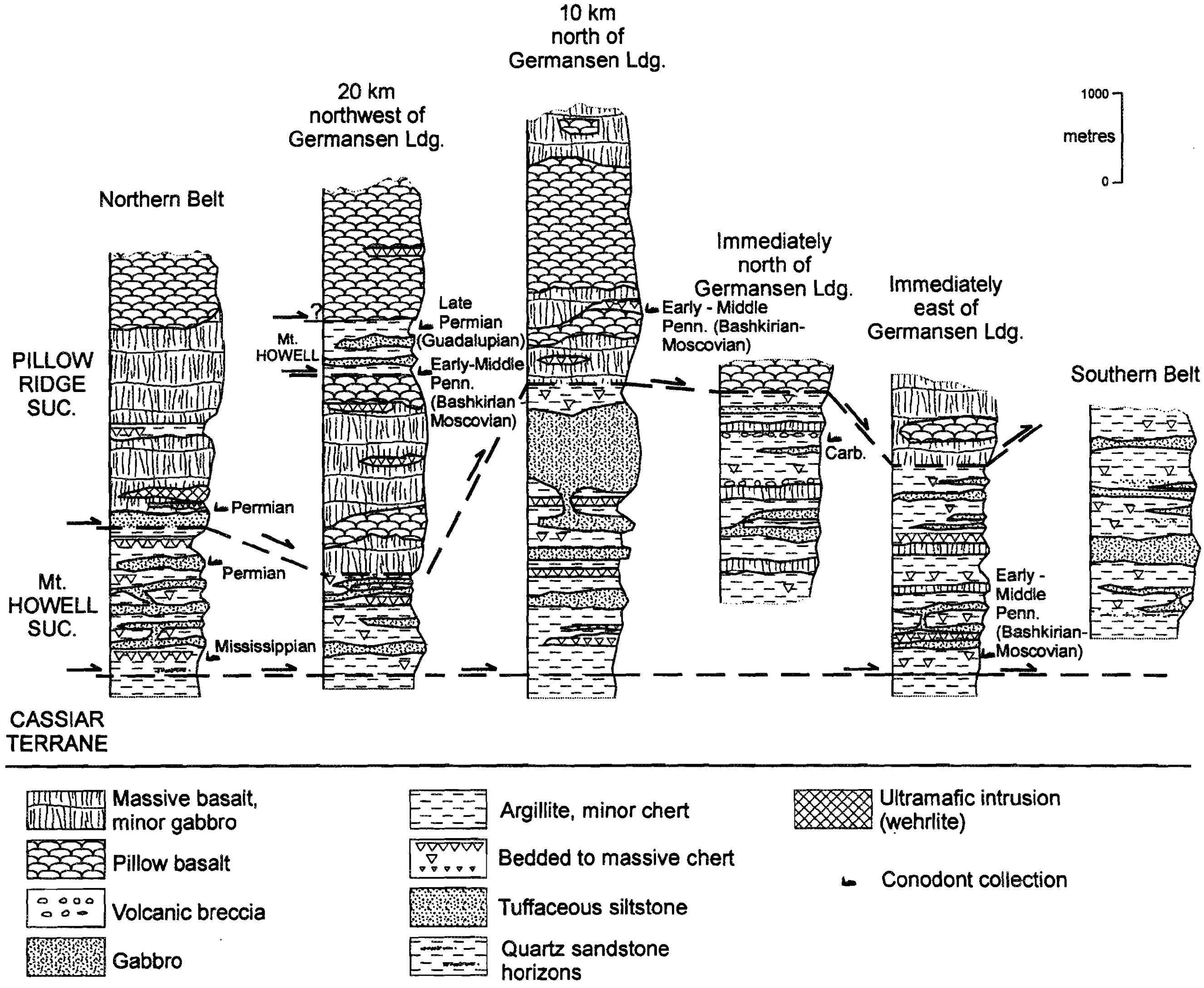


**Fig. 6.** Generalized structural–stratigraphic columns of the Nina Creek Group for selected areas of the map area. For illustrative purposes, some of the thinner units are portrayed thicker than normal.

Conodont collections from cherts within the Pillow Ridge succession are limited, but have yielded Early to Middle Pennsylvanian and Permian ages. Bashkirian to Moscovian conodonts were found near the base of the basaltic sequence north of the Omineca River, whereas conodonts of Permian age occur at the same structural level near the Osilinka River, suggesting that the thrust carrying these rocks has cut upsection. Immediately north of the Omineca River, Pillow Ridge basalts overlie Mount Howell sediments that contain Late Permian (Guadalupian) conodonts. The nature of this contact is not understood in this area. If it is conformable, at least some of the basalts would be as young as Late Permian. The presence of cryptic thrust repetitions, however, including one at the contact with the Mount Howell succession itself, cannot be ruled out. Mississippian or older basalt, on which the Mount Howell succession presumably rested, was either not deposited, suggesting the Mount Howell succession was floored by attenuated continental crust, or has not been documented.

### Mafic–Ultramafic unit

A composite body of basalt, gabbro, amphibolite, and serpentinite straddles the Osilinka valley north of Uslika Lake. It forms a fault-bounded package within the Lay Range Assemblage, but its stratigraphic relationship with this or any other unit is unclear. The age of this unit is unknown, as no fossils have been recovered from sedimentary horizons. It has closer lithologic and chemical affinities with ocean-floor igneous rocks of the Nina Creek Group than with the arc volcanics of the Lay Range Assemblage. Therefore, it is tentatively interpreted as a fault slice of the Nina Creek Group, which has been juxtaposed against the Lay Range Assemblage.

Massive to pillowed basalt forms the structurally highest and lowest parts of this package northwest of the Osilinka River. Also present are thin lenses of chert, fine- to medium-grained gabbro, and serpentinite. Mafic tuffs are associated with basalt in the lowest fault slice. Finely to very coarsely crystalline gabbro is found with serpentinite northwest and southeast of the Osilinka River. It is locally mylonitized at

**Fig. 7.** Photomicrograph of tuffaceous deposits within the Mount Howell succession. Fragments are dominated by feldspar and pyroxene, with minor chert and quartz. Field of view approximately 2.3 mm wide.

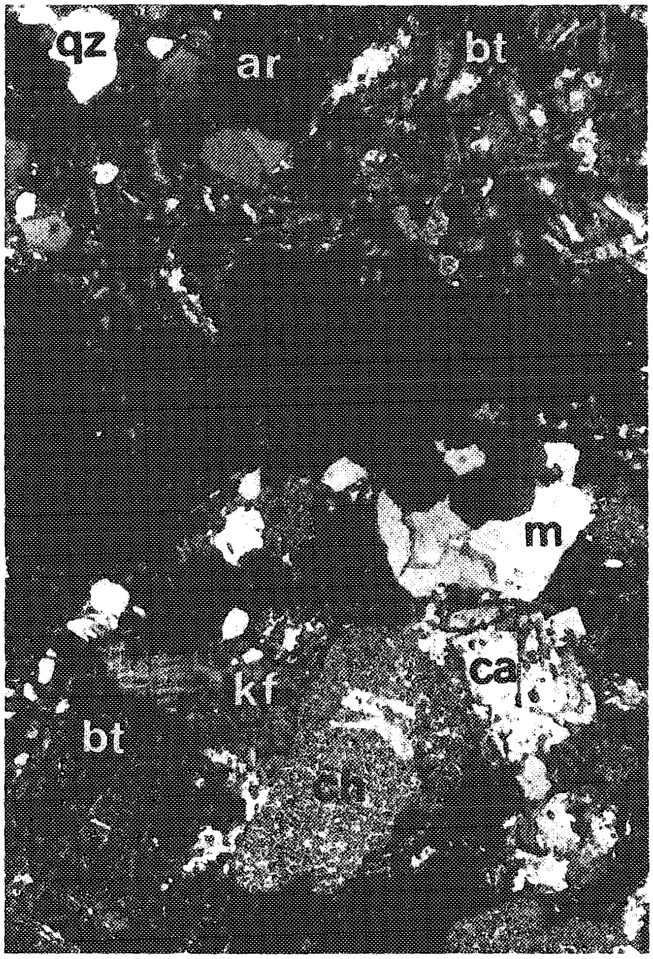
its contact with the Lay Range Assemblage. Foliated basalt, amphibolite, and gabbro make up the bulk of this unit west of the Osilinka River.

Faults bounding this unit are steep, contain transverse and oblique kinematic indicators (Ferri et al. 1992), and in one locality involve Cretaceous sediments of the Uslika Formation. This suggests that earlier structural or stratigraphic relationships between these mafic and ultramafic lithologies and other rock packages in the area have been overprinted by later strike-slip motion.

## Lay Range Assemblage

The Lay Range Assemblage has been subdivided into two informal stratigraphic units: the middle Mississippian to Middle Pennsylvanian Lower Sedimentary division composed of chert, tuff, carbonate, and clastic sedimentary rocks, followed unconformably by the Permian Upper Mafic Tuff division, which is predominantly tuff, agglomerate, and mafic volcanic flows (Ferri et al. 1993). The relationship between these divisions is visible only along the northwestern part of the Lay Range, where a large northeasterly overturned anticlinal fold has been mapped (Figs. 2, 3*b*). The bulk of the Lay Range Assemblage in the study area consists entirely of the Upper Mafic Tuff division. Its internal geometry is poorly constrained, but it generally forms steeply west dipping panels.

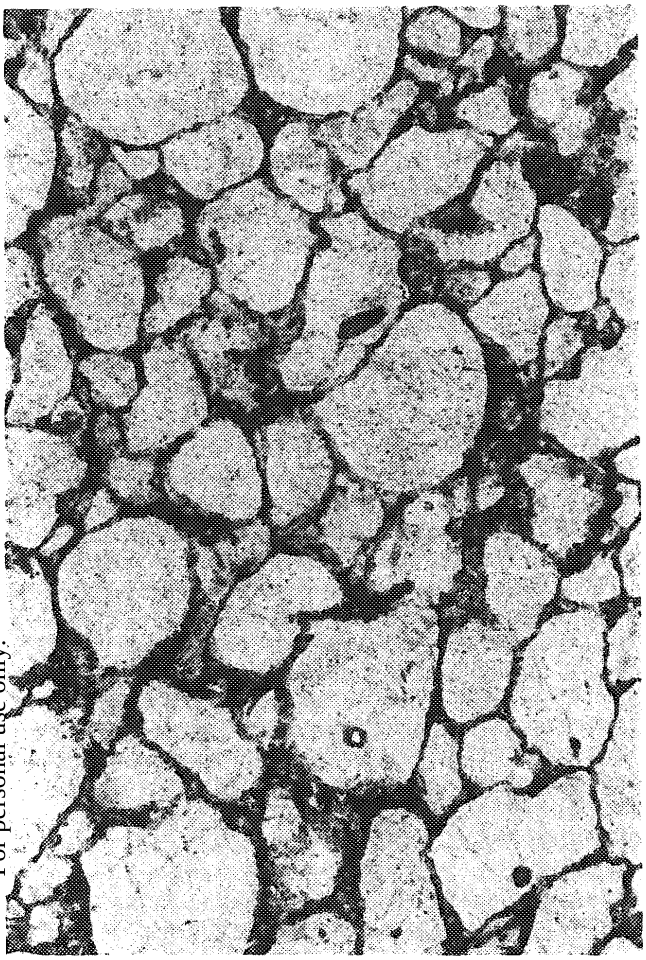
**Fig. 8.** Photomicrograph of tuffaceous deposits within the upper part of the Mount Howell succession, located approximately 5 km north of Germansen Landing. The clasts in this sample reflect a mixed volcanic–continental source. These include chert (ch), quartz (qz), carbonate (ca), argillite (ar), basalt (bt), potassic feldspar (kf), and metamorphic rock (m) fragments. Field of view approximately 2.3 mm wide.



### Lower Sedimentary division

The stratigraphy of the Lower Sedimentary division is not well defined because of structural complexity. The important features are its lithologic heterogeneity and the probable "continental" derivation of at least some of the rocks. This division consists mainly of argillite and siltstone, bedded chert, thin-bedded feldspathic and quartzitic sandstone, and chert-pebble conglomerate and "grit." The conglomerates are heterolithic, containing up to cobble-size clasts of varicoloured chert, quartz, argillite, quartzite, carbonate, tuff, and clinopyroxene-phyric volcanic rock. Less common lithologies in the division include fine- to medium-grained white

**Fig. 9.** Photomicrograph of quartz sandstone within the lower part of the Mount Howell succession. Other sandstone units within the Mount Howell succession also contain abundant chert fragments. Field of view approximately 2.3 mm wide.

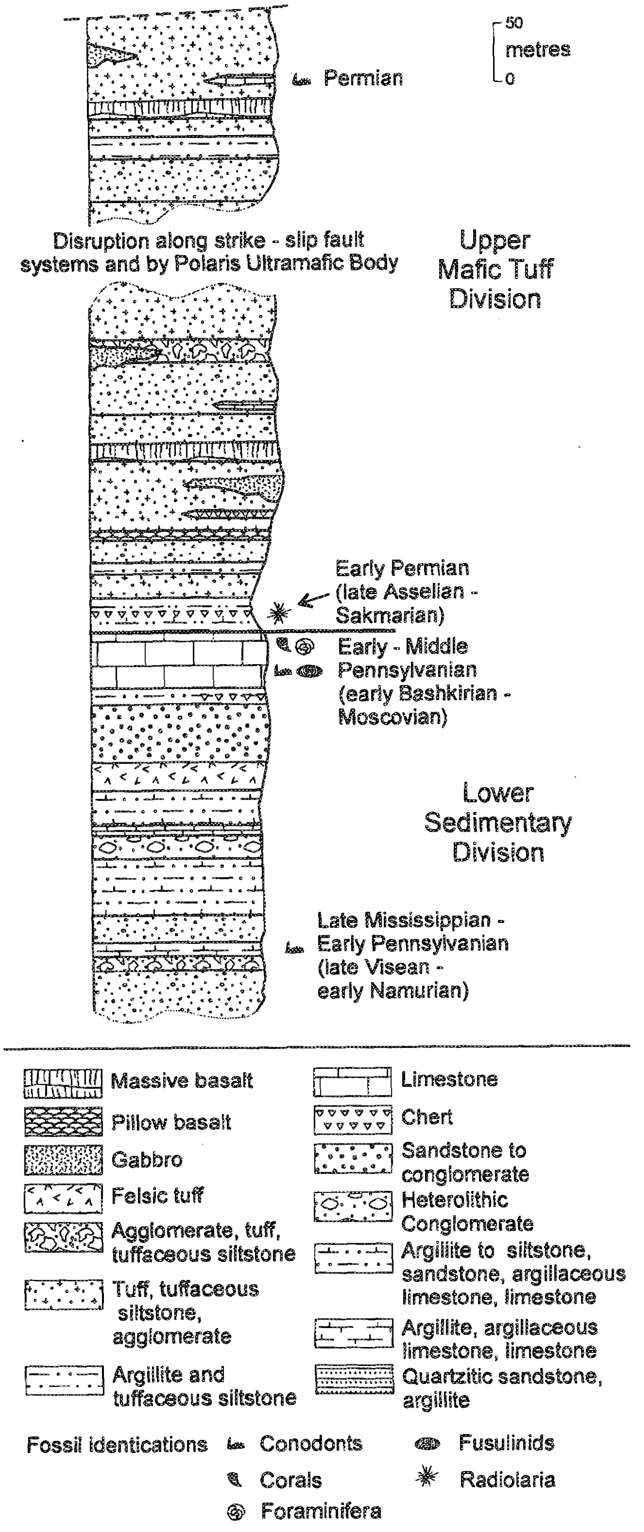


quartzite, rhyolitic tuff, shaly or thin-bedded limestone (which can be fossiliferous), limy argillite and tuffaceous rocks, and volcanic sandstones. Minor, small felsic or dioritic intrusions are present in the division, and a narrow serpentinite body is exposed along a fault zone northeast of upper Polaris Creek.

Limestone forms an important marker at the top of the Lower Sedimentary division. The largest body, in the hinge zone of the anticline, is at least 75 m thick, but elsewhere may be only a few metres thick, or absent. It is a massive to thinly bedded bioclastic limestone, locally rich in colonial and solitary horn corals, echinoderm and sponge material, fusilinaceans, and foraminifera. The limestone is locally dolomitic and contains tuffaceous layers and nodular masses of red or grey chert.

An interval, up to 50 m thick, of maroon to red, thinly bedded argillite, silty argillite, tuff, and jasperoidal chert is usually present immediately above the limestone. This intervalis succeeded by green tuffs of the upper division.

**Fig. 10.** Generalized stratigraphic column of the Lay Range Assemblage.



The known age range of the Lower Sedimentary division is from middle Mississippian to late Middle Pennsylvanian (Fig. 10). Limestone containing middle to Late Mississippian (late Visean to early Namurian) conodonts occurs near the lowest known stratigraphic level of this division; but this lower age limit is a minimum, as its base is not exposed. Fusulinaceans and foraminifera from the thick limestone at the top of the division were identified as mid-Middle Pennsylvanian (middle Moscovian; Ross and Monger 1978). Fossils collected for this study (corals, foraminifera, and fusulinaceans) are Early and Middle Pennsylvanian, ranging from early Bashkirian through early to late Moscovian (E.W. Bamber, personal communication, 1995).

### Upper Mafic Tuff division

Mafic crystal-lithic and lapilli tuffs, agglomerates, and volcanic flows are the most typical and widespread rocks of the Lay Range Assemblage. Interbedded with these lithologies are subordinate argillite, siltstone, volcanic wacke and conglomerate, chert, limy siltstone, and limestone. Volcanics of this division invariably contain phenocrysts of pyroxene and (or) feldspar up to 1 cm in length and making up to 30% of the rock. In the field, the volcanics closely resemble the Upper Triassic Takla Group, but can be distinguished by their deeper green to apple green colour due to epidote, greater induration, and stronger cleavage (where developed), and by the local but rare presence of quartz in coarser tuffs. Gabbro sills up to 100 m thick locally intrude these rocks and are traceable for several kilometres.

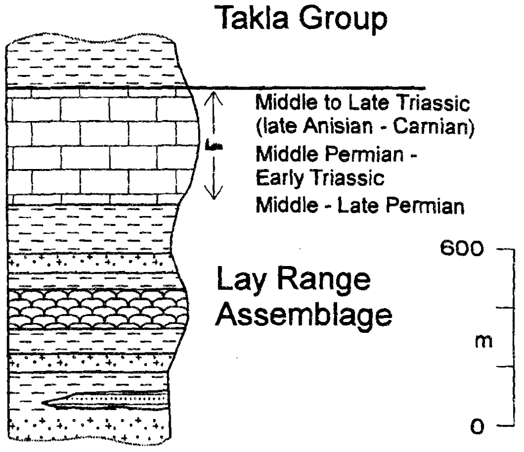
Thick sequences of bedded tuffs, tuffaceous siltstones, and lesser volcanic sandstone form the most dominant lithologies within this unit. Fragments of chert, argillite, basalt, and quartz accompany pyroxene and feldspar crystal fragments within the coarser tuffs. The coarser tuffs and lapilli tuffs may contain fragments of bryozoa, crinoid ossicles, and brachiopods. Thin, locally pillowed volcanic flows and rare conglomerate beds with argillite, chert, quartz, and volcanic clasts also occur within the tuffs. Graded, quartz-rich sandstones and wackes form a minor but conspicuous part of the tuff sequence, and are fairly common northwest of the Osilinka River.

Bedded chert near the base of the Upper Mafic Tuff division has yielded Early Permian (late Asselian to Sakmarian) radiolarians. This chert is only a few metres stratigraphically above the Early to Middle Pennsylvanian limestone of the Lower Sedimentary division. The contact between them is abrupt but otherwise unremarkable. Permian conodonts were recovered from brown-weathering, limy siltstone within mafic tuffs presumably much higher in the division, immediately west of the Osilinka River (Monger 1977; Ferri et al. 1992). Although only two fossil localities have been found within this unit, when taken in conjunction with stratigraphic relationships of the underlying Lower Sedimentary division it strongly suggests that the Upper Mafic Tuff division is entirely of Permian age.

Detrital zircons have recently been obtained from a sample of basaltic tuff from along the Osilinka River. Preliminary results for single detrital grains indicate both Middle and Late Proterozoic ages (J. Mortensen, The University of British Columbia, personal communication, 1995), indicating either the presence of underlying or adjacent continental basement or sediments of continental derivation.

Just south of the Omineca River, a section of dark green to apple green quartz-bearing tuffs, tuffaceous siltstone, augite-phyric pillow basalt, argillite, and limestone lies stratigraphically below Upper Triassic Takla Group rocks (Figs. 2, 11). The limestone at the top of the section is informally referred to as the Evans Creek Limestone. It is grey, massive, 500 m thick, and can be traced north of the river where it is truncated by several faults. The limestone contains conodont assemblages that have a combined age range of Middle Permian to Late Triassic (Fig. 11) (Ferri and Melville 1994). Conodonts from north of the river suggest a Middle to Late Triassic age (late Anisian to Carnian). South of the Omineca River, there are three conodont localities from near the top of the Evans Creek Limestone. One indicates a middle to Late Permian age, and two localities contain mixed faunal assemblages giving middle Permian and Early Triassic and middle to Late Permian and Middle Triassic ages, respectively. One locality north of the Omineca River also contains an enigmatic conodont assemblage with a single conodont of Early Triassic age mixed with others of Late Triassic (Carnian) age. Since the conodonts show no signs of reworking (M.J. Orchard, Geological Survey of Canada, personal communication, 1993), and because there is no evidence of an unconformity, significant lithological variation, or faults, the limestone appears to be a condensed sequence straddling the Permo-Triassic boundary. This is one of several localities in the Canadian Cordillera containing the stratigraphic boundary between the "basement" of Quesnellia and the succeeding Mesozoic volcanic arc (Read and Okulitch 1977).

**Fig. 11.** Generalized stratigraphic column of the Evans Creek section of the Lay Range Assemblage. Lithology patterns as in Fig. 10.



The faults bounding this sequence preclude its direct correlation with the main outcrop of the Lay Range Assemblage, but it is placed within the Upper Mafic Tuff division based on the similarities of the volcanic and sedimentary rocks, particularly the quartz-bearing tuffs.

## Structural and stratigraphic relationships between terranes

It is only in the Evans Creek area that rocks assigned to the Lay Range Assemblage apparently sit stratigraphically below Mesozoic volcanics and sediments of Quesnellia. Elsewhere contacts between terranes are marked by Late Cretaceous to early Tertiary strike-slip structures or by older, easterly directed thrust faults (Fig. 2).

Shales and slates of the Big Creek Group are succeeded by similar lithologies of the Nina Creek Group, which, in their upper part, contain gabbro sills and dikes typical of the Mount Howell succession. The overall sequence described above, together with the succeeding basalts of the Pillow Ridge succession, would suggest that the Nina Creek Group sits stratigraphically above Cassiar rocks and may represent an in situ rift succession along the edge of Ancestral North America. However, the absence of feeder dikes in underlying Cassiar rocks, together with the lack of any continental signature within Nina Creek igneous rocks (see section on Geochemistry of igneous rocks) suggests the latter have been structurally emplaced. Therefore, a cryptic thrust fault is believed to separate rocks of the Nina Creek Group from the underlying Cassiar terrane. The recessive nature of these lithologies masks their structural relationship, which is only directly demonstrated by fossil age reversals.

In the northern part of the map area, the Lay Range Assemblage sits structurally atop the Cassiar terrane. The thrust fault is readily discernible at the base of the Lay Range Assemblage, where it involves rocks of the Polaris Ultramafic Complex. Dating of synkinematic minerals from the shear-zone contact, together with kinematic indicators developed in the coarse-grained igneous rocks, indicates eastward emplacement of Lay Range rocks above the Cassiar terrane in the late Early Jurassic (Nixon et al. 1993). Locally within the cordillera, evidence suggests that rocks of the Slide Mountain and Harper Ranch terranes were imbricated and possibly thrust onto the distal edge of Ancestral North America in the Late Permian or Early Triassic, prior to the deposition of basal Mesozoic arc sequences of Quesnellia (Read and Okulitch 1977; Nelson 1993) (see Discussion).

The Lay Range Assemblage occupies the same structural position in the north as the Nina Creek Group does in the south. This suggests that, in the north, the Nina Creek Group has been structurally removed from a position between Lay Range and Cassiar rocks and thrust eastward onto the latter, where it has been subsequently removed by erosion (Figs. 4, 5).

## Geochemistry of igneous rocks

Major, trace, and rare earth element data were obtained from volcanic and intrusive rocks of the Nina Creek Group and the Lay Range Assemblage to help determine their tectonic settings (Tables 1, 2). The metamorphic grade of the sampled rocks is lower greenschist or lower, so some mobility of major elements is to be expected (Pearce 1980, 1982, 1983). The discrimination and spider diagrams presented concentrate on the more immobile and reliable trace and rare earth elements, although some major and minor elements, such as Sr, K, Rb, and Ba, have been included in some cases for comparison. More geochemical data on Nina Creek igneous rocks can be found in Ferri and Melville (1994) and Gabrielse (1975).

### Nina Creek Group

Nine analyses of the Mount Howell and Pillow Ridge successions are presented here, four of aphanitic basalt and five of gabbro (Table 1; Figs. 12–16). Based on major elements, the volcanics are tholeiitic basalts; similar results from the gabbros support the inference that they are feeders to the basalts (Table 1). Chondrite-normalized rare earth element (REE) patterns for these samples are shown in Fig. 12. These relatively flat REE patterns are typical of both island-arc tholeiites and mid-ocean-ridge basalts (MORB) (White and Patchett 1984; Jakes and Gill 1970).

Further refinement of the original tectonic setting of these basalts can be gathered from multielement distribution (spider) diagrams and discrimination diagrams utilizing Hf–Th–Ta and Ti with V, Zr, and Y (Figs. 13–16). The trace element spider diagram displays a relatively flat pattern showing no enrichment of the relatively immobile elements Th through to Cr with respect to MORB lavas (Fig. 13). The MORB setting of these igneous rocks is further supported by the discrimination diagrams in Figs. 14–16. The mid-ocean-ridge environment of the basalts and associated gabbros is clearly indicated by the immobile element covariation diagrams utilizing Ti/V and Cr/Y (Figs. 15 and 16, respectively).

Chondrite-normalized REE plots, trace-element spider diagrams, and discrimination plots for basalts of the Mafic–Ultramafic unit display the same overall patterns as those of the Nina Creek Group, suggesting that they are also MORB lavas (Figs. 12–16). Variations occur in the trace-element spider diagrams where these lavas are slightly depleted in Zr and Y.

Metamorphism of the Nina Creek Group precludes the use of the relatively mobile large ion lithophile elements (K, Rb, Ba, and Th), together with high field strength elements (Nb, Ta, Zr, Hr, and Ti; Saunders and Tarney 1984), in differentiating between back-arc basin and mid-ocean-ridge tectonic setting.

### Lay Range Assemblage

All 11 samples of the Lay Range Assemblage are from the Upper Mafic Tuff division (Table 2). Major element distributions indicate a calc-alkaline to tholeiitic affinity for these basalts. Trace and rare earth element data indicate two tectonic environments: 10 samples are consistent with arc-derived, calc-alkaline basalts and gabbro, whereas one sample contains a MORB signature (Figs. 14–18). Chondrite-normalized REE plots for the first 10 samples show overall enrichment of the elements Eu through to La that is typical of calc-alkaline, arc-related volcanism (Fig. 17) (White and Patchett 1984). This inferred tectonic setting is reinforced by the trace-element spider diagram showing relative reductions of the elements Ti to Cr with respect to MORB values and a depletion of the elements Ta and Nb relative to Ce and Th (Fig. 18) (Pearce 1983). The Hf–Ta–Th discrimination plot clearly shows the island-arc, calc-alkaline nature of these basalts (Fig. 14), and the Ti/V and Cr/Y covariation diagrams strongly suggest an island-arc setting (Figs. 15, 16). Titanium values are relatively high for these arc basalts, resulting in overlap of the data spread into the MORB field (Fig. 15). These transitional trends are not as clearly displayed by other immobile elements.

One sample of basalt from the Upper Mafic Tuff division displays MORB trends typical of basalts from the Mafic–Ultramafic unit, tentatively assigned to the Nina Creek Group (Figs. 14–18). This sample is from a poorly exposed, northeast-facing slope south of the Mesilinka River and occurs just above the inferred basal thrust carrying Lay Range rocks. It may represent a slice of the Mafic–Ultramafic unit (or the Nina Creek Group itself) incorpo-

**Table 1.** Major oxide, trace element, and rare earth element analysis for selected samples of the Nina Creek Group.

| | Pillow Ridge succession | | | | Mount Howell succession | | | | | Mafic – Ultramafic unit | |
|---|---|---|---|---|---|---|---|---|---|---|---|
| Sample No.: | FFe88-29-10 | FFe89-6-10 | FFe89-18-09 | SFD91-22-1 | FFe87-29-1 | FFe89-6-7 | FFe89-23-27 | FFe89-29-17 | SFD91-39-2 | CJR91-8-5 | FFe91-9-4 |
| Rock type: | Basalt | Basalt | Basalt | Basalt | Gabbro | Gabbro | Gabbro | Gabbro | Gabbro | Basalt | Basalt |
| UTM coordinates | | | | | | | | | | | |
| Easting: | 394341 | 385919 | 376490 | 374962 | 419877 | 386093 | 378089 | 381971 | 375390 | 362882 | 360025 |
| Northing: | 6186346 | 6203446 | 6203168 | 6209702 | 6157562 | 6203782 | 6217301 | 6215827 | 6217045 | 6220132 | 6226027 |
| $SiO_2$ (wt.%) | 49.70 | 48.76 | 48.61 | 48.37 | 47.37 | 50.23 | 48.91 | 47.74 | 49.05 | 47.74 | 48.42 |
| $TiO_2$ | 1.61 | 1.49 | 1.34 | 1.96 | 1.95 | 1.55 | 1.07 | 1.71 | 1.86 | 1.26 | 1.41 |
| $Al_2O_3$ | 13.94 | 14.04 | 14.40 | 13.69 | 15.03 | 14.32 | 13.93 | 14.10 | 15.92 | 15.64 | 13.84 |
| $Fe_2O_3$[a] | 11.66 | 10.72 | 10.63 | 12.35 | 6.09 | 10.79 | 9.74 | 11.74 | 11.20 | 10.74 | 11.74 |
| MnO | 0.21 | 0.16 | 0.17 | 0.19 | 0.12 | 0.18 | 0.20 | 0.20 | 0.18 | 0.19 | 0.22 |
| MgO | 5.91 | 6.22 | 7.20 | 7.65 | 6.56 | 6.96 | 8.47 | 7.45 | 4.54 | 8.48 | 8.38 |
| CaO | 10.78 | 12.60 | 10.34 | 9.35 | 16.90 | 7.94 | 11.04 | 10.35 | 10.25 | 8.95 | 9.19 |
| $Na_2O$ | 3.15 | 2.59 | 3.38 | 1.70 | 1.97 | 4.23 | 2.58 | 3.38 | 3.11 | 3.72 | 3.17 |
| $K_2O$ | 0.06 | 0.03 | 0.06 | 1.60 | 0.21 | 0.26 | 0.71 | 0.02 | 0.16 | 0.12 | 0.02 |
| $P_2O_5$ | 0.13 | 0.15 | 0.11 | 0.19 | 0.21 | 0.14 | 0.09 | 0.13 | 0.17 | 0.14 | 0.14 |
| FeO | 8.34 | 7.26 | 6.75 | na | na | 8.41 | 7.69 | 8.91 | na | na | na |
| LOI | 2.68 | 2.97 | 3.15 | 2.71 | 2.95 | 2.67 | 2.72 | 2.94 | 2.67 | 2.44 | 3.12 |
| Cu (ppm) | 53 | 57 | 68 | 50 | 25 | 50 | 28 | 7 | 48 | 46 | 11 |
| Pb | <5 | 6 | 4 | 1[b] | 15 | 6 | 4 | 4 | 0[b] | 1 | 4 |
| Zn | 84 | 111 | 85 | 104[b] | 44 | 77 | 92 | 82 | 91[b] | 88 | 153 |
| Ni | 42 | 76 | 90 | na | na | 72 | 122 | 85 | na | na | na |
| Mo[b] | 0.50 | 0.33 | <0.01 | 0.26 | 0.23 | 0.26 | 0.20 | 0.17 | 0.26 | 0.32 | 0.16 |
| Cr | 130 | 229 | 232 | 230[b] | 120 | 238 | 307 | 228 | 60 | 358[b] | 29[b] |
| Ba | 120 | 22 | 35 | 110[b] | 850 | 435 | 652 | 133 | 104[b] | 49 | 21 |
| Sr | 110 | 30 | 47 | 121[b] | 160 | 278 | 142 | 100 | 427[b] | 72[b] | 150[b] |
| Rb | <10 | 10 | 1 | 19[b] | 10 | 7 | 22 | 1 | 3[b] | 1[b] | <0 |
| Zr | 130 | 112 | 87 | 143 | 110 | 109 | 73 | 94 | 130 | 57 | 64 |
| Y | 34 | 36 | 31 | 39 | 46 | 39 | 32 | 31 | 43 | 23 | 24 |
| Nb | 23 | 11 | 8 | 4[b] | 10 | 7 | 9 | 9 | 4[b] | 2 | 2 |
| Th[b] | 0.28 | 0.25 | 0.19 | 0.3 | 0.17 | 0.16 | 0.10 | 0.33 | 0.2 | 0.3 | 0.2 |
| U[b] | 0.07 | 0.10 | 0.09 | 0.1 | 0.03 | 0.04 | 0.05 | 0.05 | 0.1 | 0.1 | 0.2 |
| Cs | 1 | 5 | 6 | 0.42[b] | 1 | 6 | 6 | 3 | 0.12[b] | 0.06[b] | 0.07[b] |
| Co | 42 | 33 | 32 | na | 27 | 31 | 34 | 38 | na | na | na |
| Sc | 40 | 36 | 39 | 42 | 42 | 39 | 40 | 38 | 34 | 42 | 39 |
| V | 362 | 297 | 302 | 331 | 360 | 318 | 238 | 361 | 330 | 294 | 313 |
| Bi[b] | 0.05 | 0.08 | 0.06 | 0.03 | 0.04 | 0.06 | 0.04 | 0.03 | 0.01 | <0.01 | <0.01 |
| Ti[b] | <0.01 | <0.01 | 0.02 | 0.06 | 0.02 | 0.05 | 0.08 | 0.02 | <0.01 | <0.01 | <0.01 |
| Li[b] | 5.26 | 2.13 | 3.93 | 4.38 | 13.05 | 26.61 | 7.73 | 8.29 | 4.51 | 7.39 | 9.82 |
| Sn | na | na | na | 15 | na | na | na | na | 18 | <15 | <15 |
| Ta[b] | 0.24 | 0.36 | 0.23 | 0.39 | 0.41 | 0.24 | 0.26 | 0.28 | 0.37 | 0.26 | 0.13 |
| Hf[b] | 2.57 | 2.43 | 1.95 | 3.21 | 1.26 | 1.44 | 1.20 | 1.97 | 1.90 | 1.23 | 1.94 |
| Be[b] | <0.01 | 0.62 | <0.01 | 2.15 | 0.43 | 0.97 | 1.48 | <0.01 | 2.27 | 2.37 | 4.32 |
| La[b] (ppm) | 3.57 | 3.67 | 3.53 | 5.0 | 2.68 | 4.32 | 5.39 | 4.30 | 5.40 | 3.52 | 3.30 |
| Ce[b] | 11.53 | 12.13 | 10.85 | 15.7 | 9.91 | 13.06 | 9.38 | 12.77 | 16.32 | 10.22 | 9.96 |
| Pr[b] | 1.94 | 2.06 | 1.78 | 2.6 | 1.91 | 2.14 | 2.12 | 2.04 | 2.59 | 1.64 | 1.73 |
| Nd[b] | 10.80 | 11.14 | 9.80 | 14.3 | 11.30 | 11.07 | 10.56 | 10.81 | 14.06 | 8.77 | 9.64 |
| Sm[b] | 3.72 | 3.55 | 3.26 | 4.5 | 4.04 | 3.83 | 3.18 | 3.36 | 4.52 | 2.90 | 3.32 |
| Eu[b] | 1.36 | 1.26 | 1.22 | 1.58 | 1.61 | 1.28 | 1.17 | 1.22 | 1.61 | 1.20 | 1.28 |
| Gd[b] | 5.27 | 4.91 | 4.42 | 6.33 | 6.41 | 5.27 | 4.97 | 4.60 | 6.32 | 4.01 | 4.49 |
| Tb[b] | 0.91 | 0.90 | 0.78 | 1.05 | 1.03 | 0.89 | 0.73 | 0.79 | 1.08 | 0.64 | 0.76 |
| Dy[b] | 6.18 | 5.84 | 5.03 | 6.76 | 6.73 | 5.66 | 4.69 | 5.19 | 7.11 | 4.36 | 5.02 |
| Ho[b] | 1.34 | 1.21 | 1.02 | 1.45 | 1.46 | 1.21 | 1.02 | 1.10 | 1.50 | 0.95 | 1.03 |
| Er[b] | 3.76 | 3.52 | 2.98 | 4.11 | 4.18 | 3.46 | 2.96 | 3.17 | 4.24 | 2.60 | 2.94 |
| Tm[b] | 0.55 | 0.53 | 0.44 | 0.58 | 0.59 | 0.48 | 0.45 | 0.48 | 0.58 | 0.35 | 0.41 |
| Yb[b] | 3.51 | 3.36 | 2.89 | 3.89 | 3.76 | 3.06 | 2.76 | 2.88 | 3.53 | 2.51 | 2.85 |
| Lu[b] | 0.52 | 0.47 | 0.42 | 0.53 | 0.56 | 0.44 | 0.40 | 0.43 | 0.46 | 0.34 | 0.40 |

**Notes:** All analyses, unless indicated, determined by laboratories of the British Columbia Ministry of Energy, Mines and Petroleum Resources utilizing X-ray fluorescence for major oxides and trace elements (Zr, Y, Nb, Sc, V, Ni, Cr, Ba, Sr, and Cs) and atomic absorption for trace metals (Cu, Pb, Zn, Co, Ni). na, not analyzed; LOI, loss on ignition.

[a] $Fe_2O_3$ is total iron; FeO determined by digestion titration.

[b] Analyses determined by laboratories at Memorial University of Newfoundland utilizing inductively coupled plasma mass spectroscopy.

**Table 2.** Major oxide, trace element, and rare earth element analysis for selected samples of the Lay Range Assemblage.

| | Upper Mafic Tuff division | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|
| Sample No.: | FFe92-5-5 | FFe91-8-6 | FFe91-10-7 | FFe91-11-14 | SFD91-35-7 | CRE92-14-4-2 | FFe92-4-18 | FFe92-5-13-2 | FFe92-27-1-3 | FFe92-50-13 | FFe91-35-7 |
| Rock type: | Basalt | Basalt | Basalt | Basalt | Basalt | Basalt | Basalt | Basalt | Basalt | Basalt | Gabbro |
| UTM coordinates | | | | | | | | | | | |
| Easting: | 6249400 | 355633 | 363839 | 364884 | 351227 | 6263725 | 6246550 | 6251400 | 6272800 | 6274625 | 354103 |
| Northing: | 348025 | 6231809 | 6223575 | 6219873 | 6240365 | 333900 | 348925 | 344525 | 327050 | 317825 | 6236547 |
| SiO$_2$ (wt.%) | 47.46 | 53.16 | 51.95 | 48.14 | 51.78 | 51.74 | 53.54 | 52.21 | 50.25 | 48.90 | 52.20 |
| TiO$_2$ | 1.68 | 0.76 | 1.31 | 0.95 | 1.15 | 0.98 | 1.23 | 1.19 | 1.10 | 1.28 | 1.32 |
| Al$_2$O$_3$ | 14.97 | 17.73 | 15.00 | 14.12 | 14.99 | 16.01 | 14.93 | 15.59 | 15.23 | 13.59 | 14.67 |
| Fe$_2$O$_3$$^a$ | 11.08 | 8.75 | 11.68 | 10.44 | 13.48 | 11.86 | 10.91 | 11.62 | 12.44 | 11.27 | 11.73 |
| MnO | 0.17 | 0.12 | 0.19 | 0.17 | 0.22 | 0.22 | 0.20 | 0.18 | 0.19 | 0.17 | 0.20 |
| MgO | 6.53 | 3.02 | 3.81 | 8.56 | 4.32 | 4.83 | 3.92 | 4.91 | 5.27 | 10.15 | 3.93 |
| CaO | 10.19 | 6.25 | 7.84 | 10.16 | 6.32 | 6.98 | 6.43 | 5.26 | 8.78 | 6.11 | 7.44 |
| Na$_2$O | 2.78 | 5.70 | 3.77 | 1.33 | 4.84 | 4.59 | 4.60 | 4.94 | 3.51 | 3.57 | 3.49 |
| K$_2$O | 1.08 | 0.68 | 0.28 | 2.05 | 0.29 | 0.35 | 0.91 | 1.11 | 0.58 | 0.47 | 0.62 |
| P$_2$O$_5$ | 0.16 | 0.19 | 0.23 | 0.20 | 0.16 | 0.19 | 0.24 | 0.24 | 0.17 | 0.21 | 0.22 |
| LOI | 3.93 | 3.15 | 3.01 | 3.06 | 2.25 | 1.97 | 2.66 | 2.55 | 2.44 | 4.21 | 3.47 |
| Cu (ppm) | na | 66 | 182 | 122 | 177 | na | na | na | na | na | 180 |
| Pb$^b$ | na | 2 | 3 | 2 | 2 | na | na | na | na | na | 1 |
| Zn$^b$ | na | 102 | 125 | 90 | 120 | na | na | na | na | na | 110 |
| Mo$^b$ | na | 0.90 | 0.52 | 0.40 | 1.33 | na | na | na | na | na | 1.06 |
| Cr | 294 | 59$^b$ | 67$^b$ | 328$^b$ | 75$^b$ | 30 | 27 | 40 | 41 | 430 | 68$^b$ |
| Ba$^b$ | 59 | 260 | 118 | 371 | 230 | 207 | 414 | 637 | 208 | 71 | 198 |
| Sr | 174 | 173$^b$ | 166$^b$ | 166$^b$ | 290$^b$ | 378 | 178 | 284 | 399 | 223 | 115$^b$ |
| Rb | 11 | 8$^b$ | 4$^b$ | 50$^b$ | 7$^b$ | < 10 | < 10 | < 10 | < 10 | < 10 | 9$^b$ |
| Zr | 121 | 65 | 95 | 90 | 81 | 82 | 102 | 96 | 69 | 97 | 95 |
| Y | 37 | 18 | 26 | 19 | 24 | 25 | 29 | 30 | 23 | 26 | 25 |
| Nb$^b$ | 3 | 3 | 5 | 7 | 32 | 4 | 6 | 4 | 3 | 11 | 5 |
| Th$^b$ | 0.2 | 1.3 | 1.5 | 1.7 | 1.3 | 1.4 | 1.8 | 1.4 | 0.9 | 1.2 | 1.5 |
| U | < 15 | 0.4$^b$ | 0.6$^b$ | 0.6$^b$ | 0.6$^b$ | < 15 | < 15 | < 15 | < 15 | < 15 | 0.6$^b$ |
| Cs | < 1 | 0.17$^b$ | 0.59$^b$ | 1.25$^b$ | 0.33$^b$ | < 1 | < 1 | < 1 | < 1 | < 1 | 0.49$^b$ |
| Sc | 44 | 21 | 33 | 42 | 39 | 40 | 33 | 35 | 41 | 23 | 33 |
| V | 307 | 198 | 367 | 296 | 380 | 347 | 329 | 340 | 402 | 204 | 370 |
| Bi$^b$ | na | <0.01 | <0.01 | <0.01 | <0.01 | na | na | na | na | na | 0.04 |
| Tl$^b$ | na | 0.03 | 0.03 | 0.18 | 0.05 | na | na | na | na | na | 0.04 |
| Li$^b$ | na | 8.52 | 17.11 | 21.13 | 12.04 | na | na | na | na | na | 18.81 |
| Sn | < 15 | < 15 | 16 | < 15 | 18 | < 15 | < 15 | < 15 | < 15 | 17 | < 15 |
| Ta$^b$ | 0.22 | 0.16 | 0.31 | 0.40 | 15.74 | 0.22 | 0.33 | 0.25 | 0.18 | 0.47 | 0.32 |
| Hf$^b$ | 3.14 | 2.42 | 2.36 | 2.17 | 2.55 | 2.02 | 2.67 | 2.47 | 1.82 | 2.30 | 2.33 |
| Be$^b$ | na | 4.80 | 7.72 | 4.82 | 8.83 | na | na | na | na | na | 7.62 |
| La$^b$ (ppm) | 4.35 | 9.80 | 11.08 | 13.98 | 8.57 | 9.12 | 12.60 | 10.37 | 6.33 | 10.57 | 12.36 |
| Ce$^b$ | 13.86 | 22.10 | 26.88 | 31.81 | 21.42 | 21.69 | 29.20 | 25.53 | 15.47 | 22.71 | 28.46 |
| Pr$^b$ | 2.28 | 3.11 | 3.82 | 4.24 | 3.02 | 2.99 | 3.93 | 3.71 | 2.24 | 2.93 | 3.88 |
| Nd$^b$ | 12.01 | 13.74 | 17.96 | 19.34 | 14.31 | 14.01 | 18.47 | 17.10 | 10.44 | 12.55 | 18.06 |
| Sm$^b$ | 3.95 | 3.50 | 4.45 | 4.40 | 3.72 | 3.38 | 4.39 | 4.45 | 2.84 | 3.18 | 4.38 |
| Eu$^b$ | 1.46 | 1.11 | 1.51 | 1.35 | 1.18 | 1.10 | 1.43 | 1.33 | 1.00 | 1.14 | 1.46 |
| Gd$^b$ | 5.48 | 3.54 | 4.72 | 4.76 | 4.14 | 3.80 | 4.44 | 4.80 | 3.42 | 3.86 | 4.71 |
| Tb$^b$ | 0.86 | 0.52 | 0.71 | 0.60 | 0.68 | 0.54 | 0.69 | 0.69 | 0.49 | 0.60 | 0.70 |
| Dy$^b$ | 6.49 | 3.28 | 4.51 | 3.45 | 4.27 | 3.93 | 4.56 | 4.92 | 3.40 | 4.42 | 4.31 |
| Ho$^b$ | 1.22 | 0.68 | 0.91 | 0.68 | 0.88 | 0.81 | 0.85 | 0.96 | 0.70 | 0.91 | 0.87 |
| Er$^b$ | 3.57 | 1.85 | 2.71 | 1.92 | 2.54 | 2.21 | 2.61 | 2.74 | 2.13 | 2.53 | 2.52 |
| Tm$^b$ | 0.52 | 0.29 | 0.38 | 0.25 | 0.36 | 0.31 | 0.36 | 0.41 | 0.31 | 0.35 | 0.38 |
| Yb$^b$ | 3.32 | 1.84 | 2.44 | 1.73 | 2.41 | 2.12 | 2.44 | 2.63 | 2.06 | 2.23 | 2.40 |
| Lu$^b$ | 0.50 | 0.31 | 0.34 | 0.25 | 0.35 | 0.33 | 0.37 | 0.41 | 0.32 | 0.31 | 0.36 |

**Notes:** All analyses, unless indicated, determined by laboratories of the British Columbia Ministry of Energy, Mines and Petroleum Resources utilizing X-ray fluorescence for major oxides and trace elements (Zr, Y, Nb, Sc, Vi, Ni, Cr, Ba, Sr, and Cs) and atomic absorption for trace metals (Cu, Pb, Zn, Co, Ni). na, not analyzed.

[a] Fe$_2$O$_3$ is total iron.

[b] Analyses determined by laboratories at Memorial University of Newfoundland utilizing inductively coupled plasma mass spectroscopy.

**Fig. 12.** Chondrite-normalized rare earth element distribution patterns for Nina Creek and basalts and gabbros of the Mafic–Ultramafic unit. Normalization values from Andrews and Ebihara (1982).

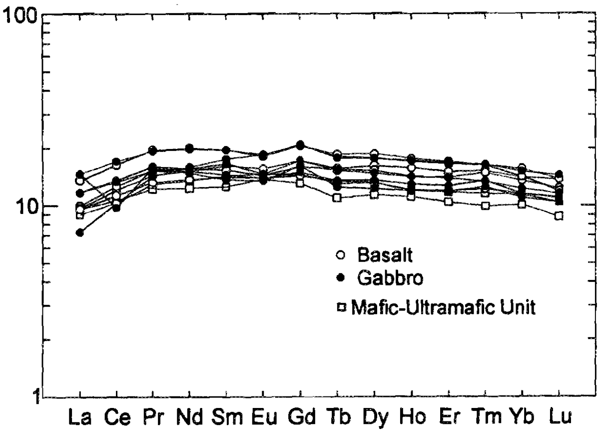
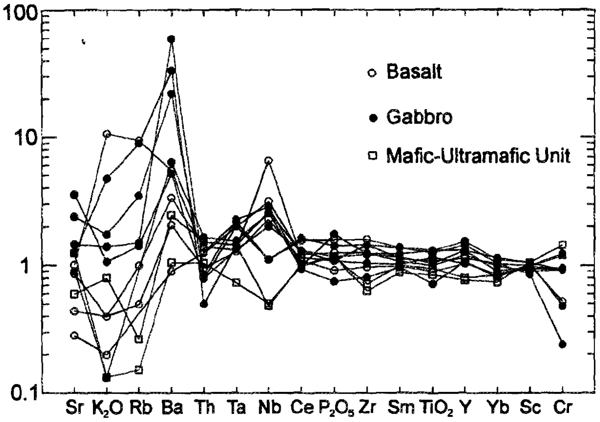


**Fig. 13.** MORB-normalized multielement distribution patterns for Nina Creek and Mafic–Ultramafic basalts and gabbros. Normalization values from Pearce (1982).



rated within this fault zone. Further work is required to determine the affinity of these basalts.

## Evidence for links between Ancestral North America, Slide Mountain, and Harper Ranch terranes

One of the most important relationships in the study area is the mixed conodont fauna within the Evans Creek Limestone, which suggests a paraconformity or condensed sequence. This links the Lay Range Assemblage and the overlying Takla Group. This adds support to the assignment of the Lay Range to the Harper Ranch subterrane, which in southern British Columbia forms part of the "basement" of Quesnellia. This is also indirectly substantiated by the Polaris Ultramafic Complex, which intrudes the Lay Range Assemblage and is believed to have been a local source for overlying Mesozoic mafic volcanics of Quesnellia (Nixon et al. 1990).

**Fig. 14.** Th–Hf–Ta tectonic discrimination diagram of Wood (1980) as applied to igneous rocks of Nina Creek Group and Lay Range Assemblage. A, N-type MORB; B, E-type MORB and tholeiitic within-plate basalts and differentiates; C, alkaline within-plate basalts and differentiates; D, destructive plate margin basalts and differentiates (the broken line separates tholeiitic island-arc basalts (upper part) from calc-alkaline basalts (lower part)).

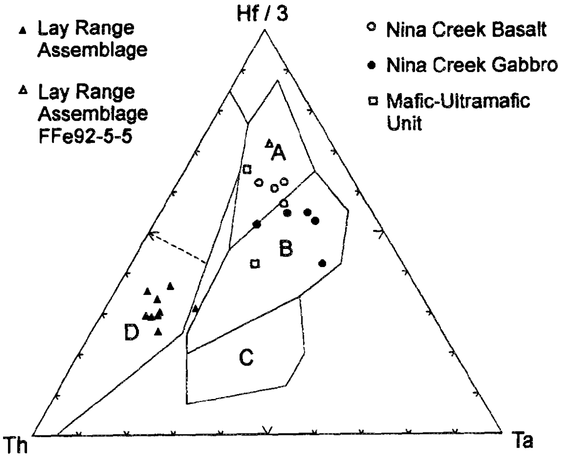
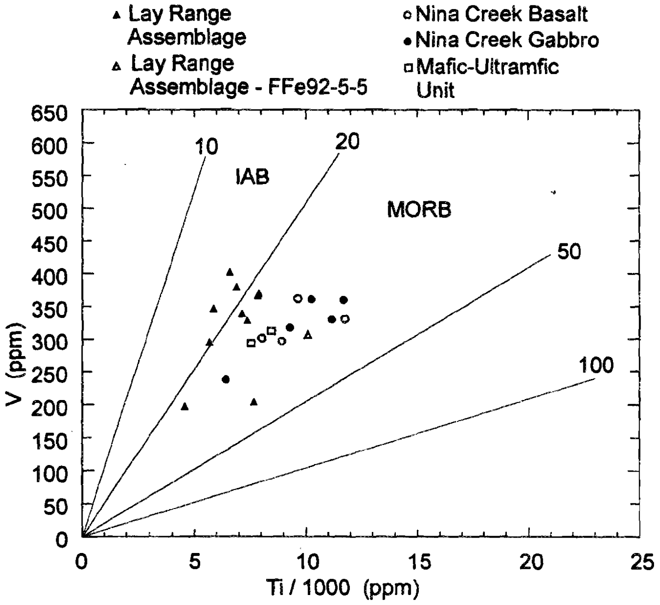


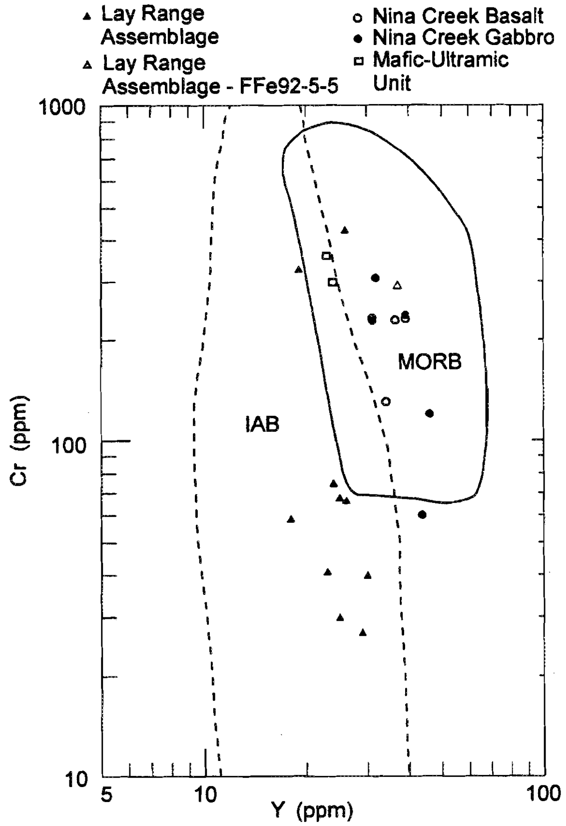
**Fig. 15.** Plot of Ti/1000 vs. V utilizing data for igneous rocks of the Nina Creek Group and Lay Range Assemblage (after Shervais 1982). IAB, island-arc basalts; MORB, mid-ocean-ridge basalts.



There are no direct stratigraphic links between the Lay Range Assemblage, the Nina Creek Group, and the Cassiar terrane, but an original continuity between them may be inferred from minor but distinctive lithological units within them.

The Mount Howell succession of the Nina Creek Group contains sections of tuffaceous rocks that contrast with the predominant argillaceous sediments and ocean-floor basalts of that unit. They appear to represent turbiditic deposits shed from a different source to the west or east. The stratigraphic position of these tuffs is uncertain. Since these subunits contain quartz and metamorphic detritus as well as feldspar- and

**Fig. 16.** Cr vs. Y discrimination plot for igneous rocks of the Nina Creek Group and Lay Range Assemblage (after Pearce 1982). IAB, island-arc basalt; MORB, mid-ocean-ridge basalt.



pyroxene-bearing tuffs (Figs. 7, 8), a continental as well as contemporary mafic volcanic provenance beyond the basin is implied. Significant mafic volcanism younger than Early Carboniferous is not documented anywhere in the Cassiar and Kootenay terranes (Gordey 1981; Nelson 1993; Gabrielse 1991; Struik 1987; Schiarizza and Preto 1987; Bamber et al. 1991; Monger et al. 1991). A more likely source of the Mount Howell tuffs is the Lay Range Assemblage of the Harper Ranch subterrane. The tuffs could thus be distal-facies equivalents of the Upper Mafic Tuff division, which they superficially resemble, with reworked material from the Lower Sedimentary division, which would account for the "continental" input.

In contrast, the lower part of the Mount Howell succession contains sandstones composed almost entirely of well-rounded quartz and (or) chert detritus (Fig. 9). This material was almost certainly derived from a piece of continental material, either the North American Margin to the east or from the Lower Sedimentary division of the Lay Range Assemblage to the west, which is believed to represent a piece of distal Ancestral North America rifted off during the initiation of the Nina Creek ocean. The absence of mafic volcanic material in sandstones of the Nina Creek Group argues against derivation from the Lay Range arc, suggesting a possible link to North America, in agreement with Gabrielse (1991) and Nelson (1993) with respect to certain Slide Mountain sandstones in the Sylvester Allochthon.

**Fig. 17.** Chondrite-normalized rare earth element distribution patterns for volcanic rocks from the Lay Range Assemblage. Normalization values from Andrews and Ebihara (1982).

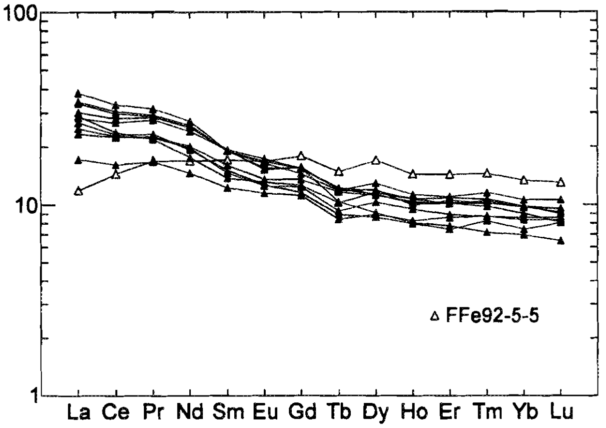
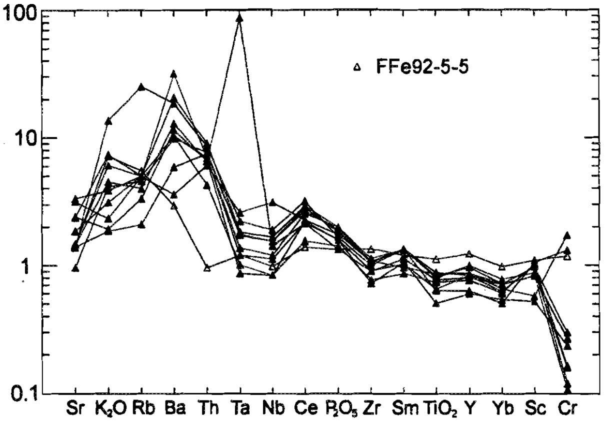


**Fig. 18.** MORB-normalized multielement distribution patterns for volcanic rocks from the Lay Range Assemblage. Normalization values from Pearce (1982).



Therefore, arguments can be made for indirect linkages between the Slide Mountain terrane and Ancestral North America in the Mississippian and for Slide Mountain and Harper Ranch terranes from the Mississippian to Permian. A direct link between the Harper Ranch and Quesnel terranes is demonstrated by the clear basement relationship of the Lay Range Assemblage to overlying Mesozoic volcanics of the Takla Group. The main uncertainties concern the maximum width of the Slide Mountain basin, and the possible role of large latitudinal displacements during formation or accretion of these terranes (Richards et al. 1993).

## Discussion

### Regional correlations

Other representatives of the Slide Mountain terrane include division II of the Sylvester Allochthon in the Cassiar Mountains (Nelson 1993; Harms 1986), the Antler Formation in the northern Cariboo Mountains (Struik and Orchard 1985), and the Fennell Formation north of Kamloops (Schiarizza and Preto 1987) (Fig. 1). These units have much in common, and their ocean-floor character has been well documented. The tectonic environment of the Harper Ranch subterrane is less well understood.

The Harper Ranch Group in the type area in southern British Columbia comprises, in part, Devonian to Pennsylvanian sedimentary and volcanic rocks (Smith 1974; Monger and McMillan 1984; Danner 1995) similar to those of the Lower Sedimentary division of the Lay Range Assemblage. It, too, contains a Pennsylvanian limestone, but it is commonly succeeded by Early Permian limestone with McCloud-type fauna, as opposed to the thick Permian volcaniclastic deposits of the Upper Mafic Tuff division. However, in the Hedley area of southern British Columbia, there are volcaniclastics that are interpreted to be basement to structurally overlying Late Triassic Nicola Group sediments (Ray and Dawson 1994). The package includes chert and quartz-bearing mafic tuffs, with lesser argillite and chert pebble conglomerate. The tuffs are virtually indistinguishable from the overlying arc volcanics of the Upper Triassic Nicola Group (Ray and Dawson 1994). A similar problem occurs in differentiating Upper Mafic Tuff division volcaniclastics from volcanics of the Upper Triassic Takla Group. These characteristics suggest that basement rocks to the Nicola Group in the Hedley area are similar to those of the upper part of the Lay Range Assemblage.

Near Trail, Pennsylvanian(?) to Early Triassic argillite, greywacke, conglomerate, limestone, and diabase of the Mount Roberts Formation have been assigned to the Harper Ranch subterrane. This is based on the presence of McCloud faunal assemblages from interbedded Permian limestones. These rocks are interpreted to have been deposited adjacent to an active volcanic arc (Roback and Walker 1995). The authors suggest the presence of continental basement below the Mount Roberts Formation through the occurrence of Precambrian detrital zircons within interbedded sandstones. They also propose that the continental source of these sediments represents a rifted piece of distal Ancestral North America.

An equivalent of the Lay Range Assemblage 350 km to the north of the study area is likely present in division III of the Sylvester Allochthon, a structural panel of arc-related igneous and sedimentary rocks (Nelson and Bradford 1993; Nelson 1993). The oldest subdivision, designated unit IIIvs, is an undated and very heterogeneous unit of maroon tuff, slate, chert, limestone, and quartz–plagioclase grit that is intruded by Early Mississippian quartz diorite and diorite. Nelson (1993) correlated this package with part of the pericratonic Yukon–Tanana terrane, based on its overall characteristics and the Early Mississippian U–Pb age and Precambrian inheritance of the diorite (Gabrielse et al. 1993). She further suggested that these rocks form the "basement" to a younger sequence in division III, namely the Pennsylvanian to Permian Huntergroup volcanics, which consist of augite–plagioclase and hornblende-phyric volcanic and volcaniclastic rocks. These are distinguished by their calc-alkaline and island-arc tholeiite geochemical signature. Although they are in part older, the Huntergroup volcanics closely resemble the Upper Mafic Tuff division of the Lay Range Assemblage, and the underlying rocks of unit IIIvs are clearly similar to those of the Lower Sedimentary division. A continental or pericratonic source terrane is indicated for the Lay Range Assemblage by the quartzofeldspathic lithologies of the lower division and by the Proterozoic detrital zircons in the Upper Mafic Tuff division. This source terrane may also form the basement to the Lay Range Assemblage. This supports Nelson's (1993) correlation between the Harper Ranch subterrane and part of the Yukon–Tanana terrane. A further similarity between the Sylvester Allochthon and the study area is that division III structurally overlies the ocean-floor division II of the allochthon (assigned to the Slide Mountain terrane by Nelson 1993), both of which were thrust onto the Cassiar terrane, a structural arrangement that applies also to the Lay Range Assemblage, the Nina Creek Group, and the Cassiar terrane, respectively.

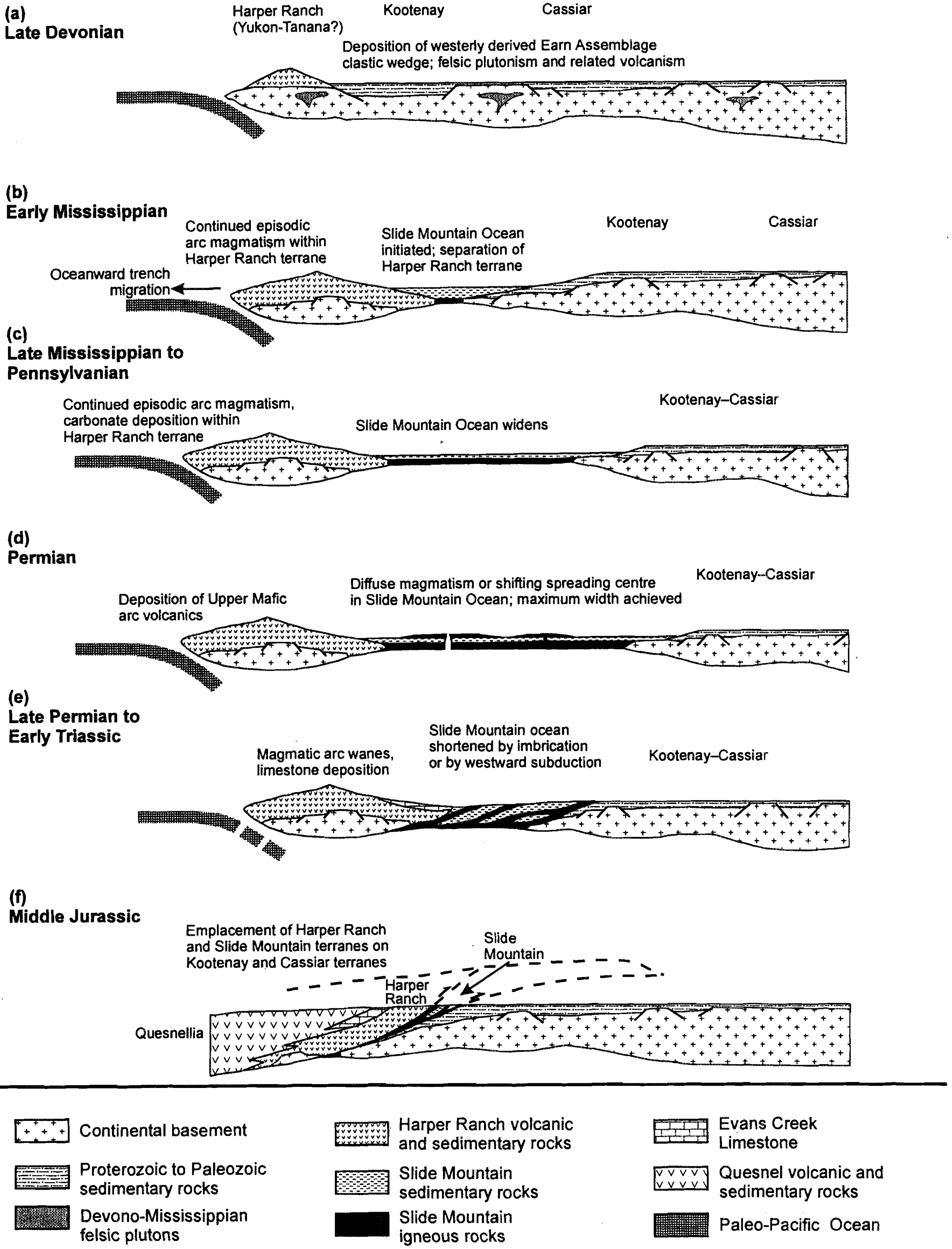
### Tectonic history

The order of structural stacking referred to above must form the framework of any tectonic model for these terranes. The structurally highest entity was presumably the most outboard during their formation in the late Paleozoic, assuming generally eastward tectonic vergence during later accretion. Based on this and the interpretation of the terranes given earlier, the simplest tectonic hypothesis is that the Lay Range Assemblage was founded on a piece of continental crust that rifted away from the North American miogeocline in the Late Devonian to Early Mississippian, forming a marginal basin represented by the Nina Creek Group or Slide Mountain terrane. The basin received sediment, at least initially, from both sides. Ocean-floor volcanism continued, at least intermittently, into the Permian, when it probably reached its maximum width. Eventually, the basin contracted during subsequent plate (or microplate) convergence events, ending with obduction onto the miogeocline in the Jurassic. This basic scenario has been put forward by many workers over the years (Davis et al. 1978; Tempelman-Kluit 1979; Hansen 1990; Gabrielse 1991; Nelson 1993; Struik 1987; Mortensen 1992; Roback et al. 1994; Roback and Walker 1995), and is supported in this paper. One of the main questions is, How much of a role did subduction and back-arc extension play in the formation of the Harper Ranch subterrane and Slide Mountain terrane? Furthermore, What was its polarity and duration, and was the oceanic basin in the order of hundreds or thousands of kilometres in width?

It is probable that rifting was initiated and sustained by west-facing subduction against the continental margin, producing the arc volcanics of the Harper Ranch subterrane, and forming the Slide Mountain basin by back-arc extension (Figs. 19*a*, 19*b*) (cf. Gabrielse 1991, his Fig. 10). The continental nature of the oldest parts of the Harper Ranch subterrane has been well documented (this paper; Nelson 1993; Roback and Walker 1995). Before rifting began, subduction may have been responsible for mid-Paleozoic deformation that was contractional in nature in the northern Yukon (Norris 1968) and in the southern Cordillera (Antler orogeny) (Silberling and Roberts 1962; Speed and Sleep 1982; Smith et al. 1993), and elsewhere was manifested by volcanism and deposition of the westerly derived chert and quartz-rich clastics of the Earn Assemblage (Gordey 1991).

North- or northeast-dipping subduction below the Yukon–Tanana terrane during Late Devonian to mid-Mississippian time has been suggested by Mortensen (1992), and Early Mississippian (to Permian?) rifting within the North American cratonic margin is well documented in the subsurface along the Peace River Arch where a series of grabens can be

**Fig. 19.** Schematic diagram (not to scale) summarizing inferred evolution of Ancestral North America, Slide Mountain, Harper Ranch, and related terranes from the late Paleozoic to the early Mesozoic.



traced into the deformed belt (Barclay et al. 1990). An early, magmatic expression of the arc-rifting event may be the distinctive Devono-Mississippian felsic intrusions, or augen orthogneisses, and felsic volcanics found locally in the older parts of the Harper Ranch subterrane, in the pericratonic Yukon – Tanana and Kootenay terranes, and in the Earn Assemblage of the Cassiar terrane. These felsic rocks were likely emplaced or deposited while these tectonic elements were still quasi-contiguous (Gordey 1981; Nelson 1993; Mortensen 1992; Mortensen et al. 1987; Schiarizza and Preto 1987). Late Devonian to Early Mississippian felsic tuff within the Big Creek Group is a reflection of this arc magmatism in the study area (Ferri and Melville 1994; J. Mortensen, The University of British Columbia, personal communication, 1995).

The eastward subduction beneath the Harper Ranch subterrane was probably episodic (Figs. 19*c*, 19*d*), since the various parts of the subterrane contain relatively restricted sections of arc volcanics, separated by sedimentary deposits. For example, in the Lay Range Assemblage, the Upper Mafic Tuff division records significant arc volcanism in the Permian, also a time when thick sections of ocean-floor basalt were being erupted in the Nina Creek Group basin to the east. Preceding this, however, no Late Pennsylvanian arc volcanics are documented by fossil or direct age evidence, although they may have existed, and the unconformity above the Middle Pennsylvanian limestone marks their uplift and erosion. The Huntergroup volcanics in division III of the Sylvester Allochthon are also of this age. Erosion of these late Paleozoic volcanics and underlying continental rocks may be the source of the lenses of the quartz-bearing mafic volcaniclastics in the Mount Howell succession. The case for still older island-arc volcanism, in the rest of the Lower Sedimentary division, and in unit IIIvs of the Sylvester Allochthon, is more tenuous. These divisions are composed largely of sedimentary rocks, and contain few volcanics of Mississippian age. Mississippian volcaniclastics are best represented in the Harper Ranch Group, where andesitic to dacitic ash was redeposited as tuffaceous turbidites in a mudstone basin (Smith 1974). The older volcanic centres of the Harper Ranch arc simply may not be preserved, or they are covered by the younger rocks of Quesnellia.

Magmatism in the Nina Creek basin was probably similarly episodic (Figs. 19*c*, 19*d*). The Nina Creek Group is not a simple oceanic assemblage with sediments overlying older MORB volcanics. Indeed, only Middle Pennsylvanian and Permian volcanism is documented, although the known range might be expanded given better fossil control. The presence of gabbro sills in the upper Mount Howell succession, which fed overlying basalts, does not evoke a typical mid-ocean ridge, and there are apparently no sheeted-dike complexes to demonstrate accretionary processes. The evidence better fits a model of episodic volcanism and some form of slow, diffuse spreading, with eruption centres developing at different times and at different places within a deep-water basin. This type of ocean-floor volcanism is also consistent with a back-arc setting (Tarney et al. 1981). Taken as a whole, basalts in other areas of the Slide Mountain terrane do encompass the entire Mississippian through Permian time span of the basin (Nelson and Bradford 1993; Nelson 1993; Struik and Orchard 1985; Harms 1986; Schiarizza and Preto 1987; Klepacki and Wheeler 1985), although this could mean that sea-floor spreading was diachronous. A more precise interpretation of Slide Mountain terrane stratigraphy is hampered by its ubiquitous structural imbrication and incomplete fossil record.

The size and configuration of the Slide Mountain basin have been the subject of much discussion, with estimates varying from Pacific-size oceans (Harms 1986; Speed 1979; Snyder and Brueckner 1985) to narrow marginal seas (Klepacki and Wheeler 1985; Schiarizza and Preto 1987; Miller et al. 1992; Nelson 1993; Roback et al. 1994). If tectonic processes were more episodic than continuous, as proposed here, the average width was probably in the order of hundreds rather than thousands of kilometres. This, of course, assumes opening and closure of the Slide Mountain ocean between Late Devonian and Early Triassic times, approximately 100 Ma.

The approximate size of the Slide Mountain ocean through time can be inferred from the comparison of faunal assemblages between the various terranes, particularly between the Harper Ranch subterrane and Ancestral North America. Locally, Pennsylvanian macrofauna in the Lower Sedimentary division indicate North American – Uralian rather than Tethyan affinities during this period (E.W. Bamber, personal communication, 1994). The rugose corals *Corwenia* sp. and possibly ?*Lithostrotion* sp. and ?*Dibunophyllum* sp. are part of a fauna with European and Russian affinities that first appeared in western and northern North America during Bashkirian (Early Pennsylvanian) time. This fauna occurs in a belt extending from the Sverdrup Basin, in the Canadian Arctic, through the Yukon Territory and Alaska and into northern British Columbia. The associated fusulinacean foraminifera are of cosmopolitan aspect in part, but also show strong affinities with coeval faunas from the northwestern Canadian mainland, the Sverdrup Basin, and the Russian Platform (E.W. Bamber, personal communication, 1994).

Although this similarity between Lay Range and North American corals implies a degree of proximity between these terranes up to the Early or Middle Pennsylvanian, the same cannot be said for the Permian, because no diagnostic fossils have been found in the Upper Mafic Tuff division. However, probable correlative rocks in part of the Sylvester Allochthon (division III, see later) contain Early Permian brachiopod and coral species of Uralian – North American affinity, together with fusulinids found in southern Ancestral North America and within Cordilleran suspect terranes correlative with the Harper Ranch subterrane (Nelson 1993).

A somewhat different view of the Permian arises from data on taxa in the Harper Ranch Group and equivalent packages in the southern Cordillera. These have been assigned to the McCloud fauna or McCloud Belt, which exhibits Tethyan, endemic, and North American characteristics (Miller et al. 1992; Miller and Wright 1987; Stevens et al. 1990; Nelson and Nelson 1985). McCloud faunas suggest an environment remote from the North American realm, but with influences from both the Tethyan and North American faunal domains. Theories on the amount of separation from the continent range from several hundred to thousands of kilometres (Miller 1987; Saleeby 1983; Stevens et al. 1990). Recent work has shown that the eastern limit of the Permian Tethyan coral realm may have been approximately 1500–

2000 km off the western margin of Ancestral North America (Belasky and Runnegar 1994), allowing for the mixing of faunas, at least by tectonic transport. If valid, this makes the narrow basin option acceptable, although it does not discriminate against a much wider basin.

In summary, local faunal data and sedimentalogic links suggest the Lay Range Assemblage was possibly situated fairly close to the Ancestral North American margin until the Middle Pennsylvanian, but may have been more remote during the Early Permian when it was far enough away to have acquired more exotic faunal characteristics. This implies that the intervening Slide Mountain basin widened during this interval (Fig. 19*d*).

Some contraction of the Slide Mountain basin appears to have occurred in Late Permian to Early Triassic time (Fig. 19*e*) (Klepacki and Wheeler 1985; Gabrielse 1991), a tectonic event possibly equivalent to the Sonoman orogeny in the western United States (Silberling 1973). This orogeny refers to the deformation of the Havallah sequence within the Golconda allochthon, a possible southern equivalent of the Slide Mountain terrane. Although the contraction of the Slide Mountain ocean may have been accommodated by internal imbrication and lithospheric thickening where the terrane was relatively narrow, shortening of a marginal basin that may have exceeded 1000 km in width must have occurred by subduction.

Direct evidence for subduction in the Late Permian is found only in blueschist and eclogite rocks in the Anvil Allocththon in southern Yukon (Erdmer 1987; Erdmer and Armstrong 1988), leading these authors and Mortensen (1992) to suggest westerly subduction of Slide Mountain oceanic crust below an east-facing arc floored by the Yukon–Tanana terrane. This would have required a flip in the polarity of subduction below the Harper Ranch subterrane in the Permian. This reversal and shortening may have happened in the study area too, although there is no arc activity in the Lay Range Assemblage attributed to it. The Upper Mafic Tuff division is Early Permian and probably coeval with MORB volcanism in the Nina Creek Group, which is unlikely to have been generated in a basin undergoing contraction and subduction. The change from eastward to westward polarity, if it took place, must have occurred afterwards, and did not generate any known arc volcanics. Their absence, and that of high-pressure assemblages elsewhere in the Cordillera, suggests that, in general, Permo-Triassic shortening of the Slide Mountain basin was small enough to have occurred without substantial subduction.

Eastward subduction beneath the Harper Ranch subterrane was reestablished by the Middle Triassic (if interrupted at all), leading to the formation of the Nicola–Takla volcanic-arc assemblage of Quesnellia (Monger et al. 1991). Subduction ended in the late Early Jurassic, followed by the final collapse of the Slide Mountain basin and its obduction onto the continental margin (Fig. 19*f*) (Nixon et al. 1993). In the northern part of the study area, Nina Creek rocks are missing between the Lay Range Assemblage and the Cassiar terrane. In this area, Quesnellia is assumed to have overridden much of the miogeocline, including nearly all the Kootenay terrane, and the Slide Mountain was completely obducted to a structural level above the present erosional surface.

The possibility of a large latitudinal translation during and (or) after terrane accretion is an outstanding issue. Paleomagnetic data on Middle Pennsylvanian to Early Permian red cherts from the Sylvester Allochthon and the Antler Formation of the Slide Mountain Group indicate deposition at latitudes approximately 20° (about 2000 km) south of their present position relative to the craton (Richards et al. 1993). This would be easier to accommodate by oblique subduction of a wide ocean floor and subsequent transcurrent motion of suspect terranes along the continental margin than by closure of a marginal basin, but it would be hard to explain the North American influence in the McCloud fauna of a therefore far-traveled Harper Ranch arc. An appealing compromise is provided by Nelson (1993), who envisaged the Slide Mountain as a mostly narrow, back-arc basin, centred by a dextral transform fault system punctuated by short ridge segments. In this scenario, significant northward translation of the arc and western half of the basin is accomplished without significant separation from North America.

## Conclusions

Stratigraphic, structural, and geochemical data indicate that the Lay Range Assemblage of the Harper Ranch subterrane and the Nina Creek Group of the Slide Mountain terrane represent an upper Paleozoic island arc and adjacent, coeval marginal basin, respectively. They are genetically related, the basin forming by back-arc extension east of a west-facing arc. The arc may be built on a piece of Precambrian continental crust that rifted away from distal North America in Devono-Mississippian time. The stratigraphic record of the upper Paleozoic terranes indicates that arc activity and ocean-floor volcanism were episodic rather than continuous. The Slide Mountain basin developed by diffuse spreading, reaching a maximum width in the Permian of hundreds rather than thousands of kilometres. It partly closed in Late Permian to Early Triassic times, either by imbrication or by a shortened episode of westerly subduction that did not result in arc volcanism, and fully closed during the Early Jurassic, when both terranes were emplaced on North America.

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