A combined petrographical-geochemical provenance study of the Newland Formation, Mid-Proterozoic of Montana

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Abstract - A provenance study was conducted on the Mid-Proterozoic Newland Formation, in which petrographical features of sandstones and geochemical characteristics of shales were integrated to arrive at an internally consistent interpretation.

Sandstones of the Newland Formation are typically arkosic sands and arkoses with very well rounded quartz and feldspar grains and only minor amounts of extrabasinal rock fragments. The predominant feldspar types are K-spar and microcline, feldspar grains are smaller than quartz grains, and feldspars show little alteration due to weathering. Detrital modes of Newland sandstones (QFL diagrams) indicate that they were derived from a stable cratonic source. These petrographical features imply a source area dominated by granites and granitoid gneisses, semi-arid to arid climate, tectonic quiescence and overall peneplain conditions.

Shales of the Newland Formation are dominated by illite, quartz silt, and fine crystalline dolomite. They have small La/Th rations, relatively large Hf contents, and small contents of Cr, Co, and Ni, all indicative of derivation from crust of granitic composition. Small TiO2/AI2O3 ratios also suggest source rocks of granitic composition. The average chemical index of alteration (CIA) for Newland shales is 71.8, which in light of the probable granitoid source indicates modest amounts of chemical weathering. Relatively large Si02 contents and large K20/Na20 ratios reflect derivation from stable cratonic areas and tectonic quiescence.

Thus, in general, the petrography of sandstones and geochemistry of shales provides the same provenance clues for the Newland Formation. One notable discrepancy between the two approaches is that the sandstones indicate an arid to semi-arid climate with very minor chemical weathering, whereas the CIA of the shales indicates at least modest amounts of chemical weathering. This indicates on one hand the need to better calibrate the CIA with a large variety of muds from modern climatic settings, and on the other hand the possibility that this discrepancy is due to transport segregation.

1. Introduction

The term provenance has been used by sedimentary petrographers to encompass all the factors relating to the production of a sediment or sedimentary rock. The most important aspects of a provenance study are the identification of source rocks, relief and climate in the source area, tectonic setting, transport history, and diagenetic modifications.

As far as sandstones are concerned considerable efforts have been made to extract provenance information from compositional and textural features of sandstones, and a thorough review of the subject is given by Pettijohn, Potter & Siever (1987). Standard petrographical approaches to the identification of source rocks of sandstones are investigations of undulosity and polycrystallinity of quartz grains (Basu et al. 1975), types of feldspar present (Pittman, 1970), and rock fragments (Pettijohn, Potter & Siever, 1987). Relief and climate of the source area can be inferred from grain roundness and average degree of feldspar alteration (Folk, 1980). Tectonic setting can be determined from the relative proportion of quartz, feldspar and rock fragments (Dickinson, 1985). When applied with caution, clues to the transport history of a sandstone can come from the examination of roundness and sphericity of grains, and textural and mineralogical maturity (Pettijohn, Potter & Siever, 1987). Because diagenetic processes can cause considerable post-depositional modification of the original mineralogical composition of a sandstone (e.g. McBride , 1985, 1987; Pettijohn, Potter & Siever, 1987), every effort has to be made to identify altered and replaced mineral grains and rock fragments in order to reconstruct the original composition.

Provenance studies of sandstones have clearly reached a significant level of sophistication and a lot can be accomplished by careful examination of sandstone particles in petrographical thin sections. With regard to shales on the other hand, petrographical examination of single particles is much more time-consuming and far less illuminating because of diagenetic modification of clay minerals, the main constituent of shales. However, some progress has been made in recent years to use bulk properties of shales, in particular chemical composition, for the purpose of provenance studies. Source rocks have, for example, been identified by use of rare earth element distributions (Taylor & McLennan, 1985), Ti/Al ratios (McLennan, Fryer & Young, 1979) and La/Th v. Hf diagrams (Bhatia & Taylor, 1981). Weathering conditions and climate have been inferred by application of a variety of chemical indexes based on abundances of major elements (Englund & Jorgensen, 1973; Björlykke, 1974; Nesbitt & Young, 1982). The chemical index of alteration (CIA) developed by Nesbitt & Young (1982) appears to be particularly useful in that respect. The Si02 content and the K20/Na20 ratio of mudstones have been used successfully to determine tectonic setting of mudstones (Roser & Korsch, 1986).

In many published studies of terrigenous sediments, provenance information is typically extracted from sandstones, primarily because the technique is comparatively straightforward and only requires use of a petrographical microscope. Because of the comparably large cost and effort required for geochemical provenance studies of shales, such studies are typically only conducted when sandstones in the sequence are rare or absent, or when diagenetic or metamorphic overprint have obscured or obliterated provenance clues in sandstones.

During an in-depth study of the Newland Formation that covered aspects of stratigraphy, sedimentology, geochemistry and basin evolution (J. Schieber, unpub. Ph.D. diss., Univ. Oregon, Eugene, 1985), a large database on the geochemical composition of shales in that formation was acquired. In this report geochemically derived provenance information from Newland Formation shales is compared with recently acquired data from petrographical studies of laterally associated and interbedded sandstones, in order to assess agreement between the two approaches and to refine knowledge of Newland Formation provenance.

2. Geological setting

The Newland Formation of the Mid-Proterozoic Belt Supergroup occurs in the Helena embayment, an eastern extension of the Belt basin (Fig. 1). Its lower member is dominated by grey dolomitic shales, whereas the upper member consists of alternating shale units (some 10 to over 100 m thick) and carbonate units (10-70 m thick). No sandstones occur in the lower member of the Newland Formation, but shales with sandstones interbeds mark the transition between the lower and upper members of the Newland Formation and have informally been referred to as the 'Newland Transition Zone' (Schieber, 1987). Sedimentary structures (hummocky cross-stratification) indicate that most of these sandstones were carried into the basin by storm-induced currents, and that sandstones in the southern Big Belt Mountains were derived from the south, whereas sandstones in the northern Big Belt Mountains and Little Belt Mountains were derived from the north (Schieber, 1987). A very minor proportion of sandstone interbeds occurs throughout the upper member of the Newland Formation. With deposition of the upper member of the Newland Formation the Helena embayment probably developed into an east-west trending half graben (Schieber, 1990). The southern margin of the half graben was a syndepositional fault, north of which coarse clastic sediments of the LaHood Formation (McMannis, 1963) were deposited. A brief overview of the sedimentary history of the Helena embayment is given in Schieber (1986).

3. Sandstone petrography

Petrographical thin sections were cut from a total of 55 sandstones samples and were examined for modal composition and a variety of other petrographical features (Table 1). Twenty-eight of these samples are from the Newland Formation, 15 from the laterally equivalent LaHood Formation, and 12 from the lowermost Greyson Formation . The Greyson Formation overlies the Newland Formation (Fig. 1), and its basal sandstones are lateral equivalents of the LaHood Formation. Samples from the Greyson and LaHood formations were studied for comparison purposes. For each thin section 300 points were counted using the Gazzi-Dickinson method (Ingersoll et al. 1984).

Sandstones of the Newland and Greyson formations are primarily coarse-grained, contain the same types of detrital grains, and show considerable compositional variability (Table 1), ranging from arkose to litharenite in the Newland Formation and from subarkose to litharenite in the Greyson Formation (classification after Folk, 1980). Sand grains of all types are typically rounded to well rounded (Powers, 1953) in both formations. Sorting (Folk, 1968) is best in the Newland Formation (dominated by well to very-well-sorted sands), and intermediate in the Greyson Formation (mostly moderately sorted). The quartzose grain fraction consists predominantly of monocrystalline quartz with weak or absent undulatory extinction. Polycrystalline quartz occurs in two varieties: (a) grains with a polygonal fabric of interlocking grains and (b) grains with elongate, lenticular, interlocking, sutured crystals (Fig. 2 a). Using the Gazzi-Dickinson point-counting method causes a considerable portion of the polycrystalline quartz grains of type (a) to be counted as monocrystalline quartz. Chert only occurs in trace amounts. The feldspar fraction is composed of potassium feldspars, microcline, and albite-rich plagioclase (in descending order of abundance). Potassium feldspars are commonly clouded with alteration products and may also show microperthitic intergrowth with plagioclase (albite dominant). Feldspar grains may also show rounded inclusions of quartz (relatively common; Fig. 2b) and granophyric or micrographic

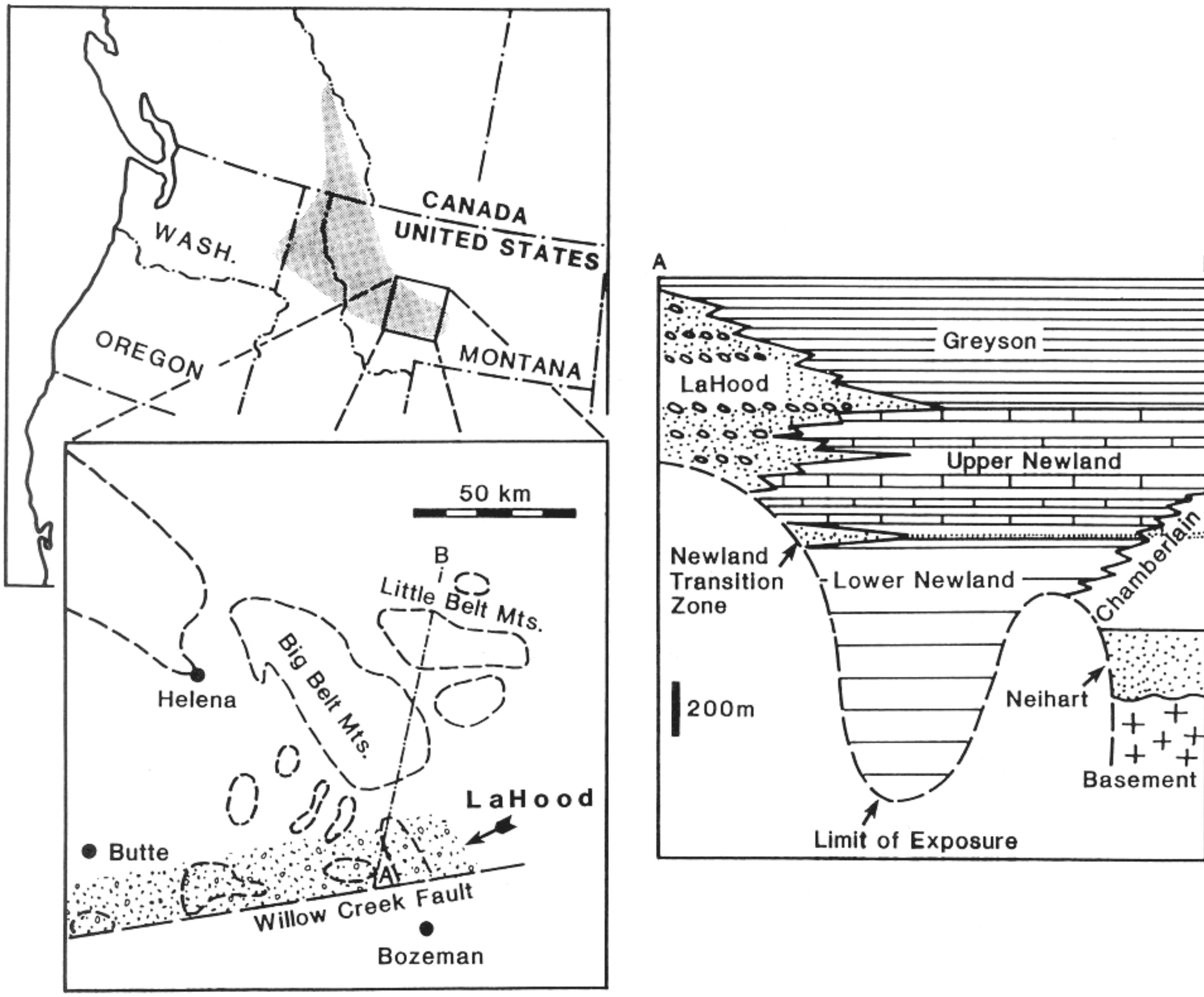

Figure 1. Location map and stratigraphical overview. Stipple pattern indicates approximate outline of Belt basin. Enlarged map shows Helena embayment portion of Belt basin. The stratigraphical overview is based on data from McMannis *(1963),* R. L. Boyce (unpub. Ph.D. diss., Univ. Texas, Austin, *1975),* and J. Schieber (unpub. Ph.D. diss., Univ. Oregon, Eugene, *1985).* It represents a generalized restored cross-section along line AB in the enlarged map portion. Areas enclosed by dashed lines indicate outcrop areas of Belt Supergroup sediments. The location of the LaHood outcrop belt is indicated by gravel pattern along the north side of the

intergrowth with quartz (Fig. 2c). Carbonate and shale particles constitute the largest proportion of rock fragments in the Newland and Greyson formation. Sedimentary rock fragments in the Newland Formation are derived from shales and carbonates of the Newland Formation itself (Fig 2d-f) and are therefore intrabasinal as well as intraformational. In the Greyson Formation a large proportion of the sedimentary rock fragments can be identified as coming from the underlying Newland Formation, with the remainder coming from the Greyson and LaHood formation. Most samples also contain a small proportion of metamorphic rock fragments (qua rtzofeldspathic gneiss, quartz muscovite schist). Fragments of quartzofeldspathic gneiss are typically counted as either monocrystalline quartz or as feldspar using the Gazzi-Dickinson point-counting method. In addition, sandstones of the Greyson Formation may contain arkose fragments that have chloritic cement (derived from the LaHood Formation).

Sandstones of the LaHood Formation are predominantly arkoses that show moderate-to-poor sorting and predominantly angular- to-subangular grains. However, in a few samples rounded to wellrounded grains of quartz and feldspar are abundant. The quartzose and the feldspar grain fractions show the same grain types as observed in the Newland and Greyson formations. In contrast to the latter two formations, the LaHood Formation shows a larger proportion and greater variety of metamorphic rock fragments, and in most samples only a small proportion of sedimentary rock fragments. Metamorphic rock fragments observed in the LaHood Formation include chlorite schist, quartz biotite schist, marble, quartz muscovite schist, amphibolite, quartz garnet muscovite schist, quartzite and quartzofeldspathic gneiss. The latter constitutes the most abundant metamorphic rock fragment type in the LaHood Formation and is also the most common metamorphic rock fragment type in the Newland and Greyson formations.

Table 1. Textural and compositional data

Sample no.	Average grain size (mm)	Sorting*	Quartz rounding (%)†	Intrabasinal rock fragments							
				Monocrystalline quartz	Polycrystalline quartz	Chert	Feldspar	Carbonate	Shale/ Siltstone	Metamorphic rock fragments	Mica
					Upper Newland Formation						
7/8/81-21	0.60	vw	80	125	4	5	32	96	38	—	—
8/7/81-7	0.45	w	76	167	7	2	95	11	18	—	—
8/11/81/7	0.70	w	74	154	4	—	44	75	23	—	—
8/12/81-7	0.35	vw	100	72	5	—	49	128	46	—	—
8/17/81-20	0.50	vw	81	152	3	2	7	120	16	—	—
8/17/81-24	0.50	vw	80	148	6	—	21	116	9	—	—
8/18/81-2	0.45	vw	95	155	1	2	139	2	1	—	—
9/1/81-17	0.50	vw	100	143	7	4	8	120	18	—	—
9/7/81-19	0.90	w–vw	100	246	18	—	36	—	—	—	—
9/7/81-20	0.90	w–vw	100	238	10	—	52	—	—	—	—
9/8/81-7	0.80	w–vw	96	214	12	—	18	2	52	2	—
7/5/82-1B‡	0.60	vw	75	182	4	2	26	80	5	—	1
7/5/82-6‡	0.80	w	80	170	9	4	68	34	2	7	6
7/5/82-16A‡	0.90	w–vw	91	154	10	8	15	101	10	2	—
7/5/82-20‡	0.60	w	84	201	6	10	32	48	—	—	3
7/6/82-4‡	0.55	w–vw	74	141	10	20	44	64	19	1	1
7/6/82-17‡	0.80	w	86	110	11	22	52	53	50	—	2
7/24/82-2	0.55	w–vw	79	168	4	4	26	90	8	—	—
7/29/82-6	0.35	w–vw	76	130	3	2	25	98	42	—	—
8/4/82-3	0.35	vw	100	145	4	6	40	92	13	—	—
8/9/82-2	0.75	w–vw	100	228	13	—	19	39	1	—	—
7/22/81-8	0.50	vw	91	159	5	1	89	32	12	1	—
7/10/82-28‡	0.40	p	79	196	15	5	14	41	25	3	1
7/11/82-5‡	1.00	w–vw	100	238	23	—	19	8	11	—	1
7/11/82-27‡	1.10	m	89	227	30	9	16	15	3	—	—
8/16/82-8	0.45	p	80	223	17	—	25	26	9	—	—
8/28/82-5	0.60	w–vw	90	245	15	—	17	17	5	1	—
8/28/82-9	0.90	m	91	138	4	—	105	38	15	—	—

					Lower Greyson Formation						
9/8/81–18	1.10	p–m	100	66	9	—	20	95	110	—	—
9/8/81-19	0.65	vw	100	223	6	—	71	—	—	—	—
9/8/81-20	0.65	w–vw	100	197	12	1	51	1	38	—	—
7/12/82-4	0.65	w–vw	86	208	19	2	8	59	3	1	—
7/12/82-10	0.55	m	81	214	18	2	55	3	6	2	—
7/12/82-12	1.20	p	84	164	15	7	30	—	81	5	—
7/13/82-14	0.80	m	80	221	9	5	39	12	14	—	—
7/13/82-18	0.55	p	74	231	13	—	49	—	3	4	—
7/24/82-25	0.75	p	76	230	8	3	40	13	2	2	2
8/5/82-14	0.65	m	n.d.	263	12	—	23	—	—	—	2
8/25/82-22	1.00	m	75	227	6	—	61	—	4	2	—
8/29/82-2	1.20	w	91	239	22	—	4	28	7	—	—
					LaHood Formation						
8/17/82-4	1.40	vp	0	119	8	—	132	2	2	2	35
8/18/82-5	1.10	p	10	191	11	—	123	—	—	5	52
8/19/82-3	0.60	p	30	248	17	—	17	—	—	3	15
8/20/82–3	0.40	w	65	247	16	1	23	—	—	3	—
8/20/82-4	0.50	m–p	70	163	9	2	25	14	84	2	1
7/22/89-1	1.10	p	0	121	5	—	114	—	3	20	37
7/22-89-3	0.70	p	0	104	3	1	137	—	—	6	49
7/22/89-4	1.00	p	0	98	4	—	136	—	1	20	41
7/22/89-5	0.90	p	0	109	3	—	141	—	1	14	32
7/22/89-6	0.90	m	0	122	1	—	138	1	—	6	32
7/22/89-7	0.90	p	10	101	2	—	143	—	—	15	39
7/22/89-9	0.75	m	5	105	4	—	162	—	1	4	24
7/22/89-10	0.35	m	5	105	0	—	136	—	—	4	55
7/22/89-16	1.25	p	0	156	4	—	102	—	2	2	34
7/22/89-18	1.25	p	0	138	2	—	134	—	1	2	23

‡ Samples from southern Big Belt Mountains.
* Sorting: vp = very poorly sorted; p = poorly sorted; m = moderately sorted; w = well sorted; vw = very well sorted.
† Quartz rounding (%): figure indicates proportion of well-rounded quartz grains.

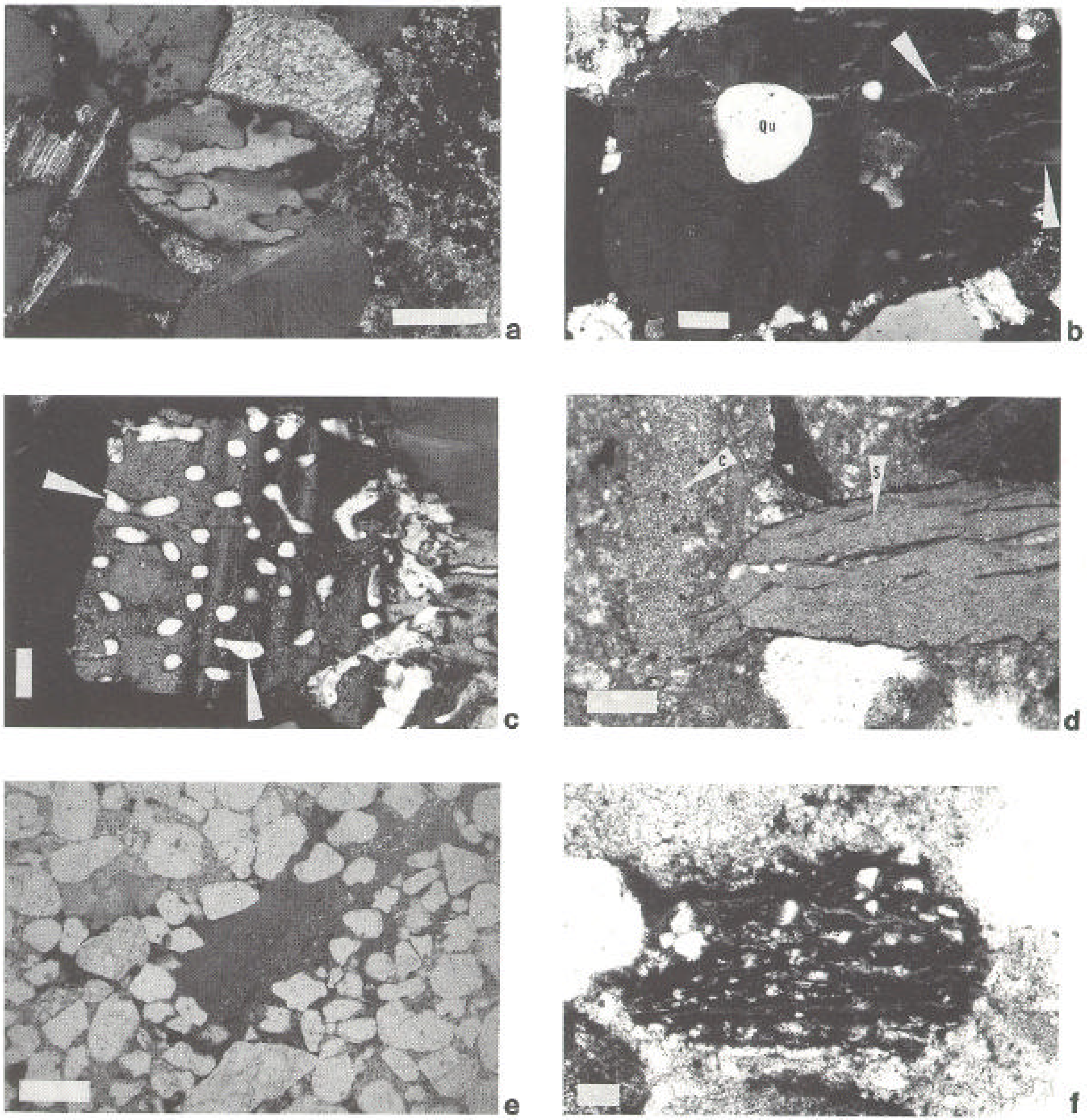

Figure 2 (a) Polycrystalline quartz grain of metamorphic origin (sample no. 8/7/81-7). Note elongate, lenticular, interlocking, and sutured crystals. Scale bar 100 ,um (crossed polarizers). (b) Grain of potassium feldspar (sample no. 8/29/82-9) that shows rounded quartz inclusion (marked as `Qu') and perthitic unmixing lamellae (arrows). Scale bar 100 ìm (crossed polarizers). (c) Grain of twinned albite with granophyric intergrowth (myrmekite) of quartz (sample no. 9/7/81-19). Quartz intergrowths are white to light grey and have a wormy-tubular morphology (arrows). Scale bar 100 ìm (crossed polarizers). (d) Intrabasinal sedimentary rock fragments (sample no. 9/8/81-18). The shale fragment (arrow S) can be identified as carbonaceous swirl shale as found in the Newland Formation (Schieber, 1989). The carbonate fragment (arrow C) with its characteristic texture consisting of equidimensional clear calcite grains was derived from `molar tooth carbonates' of the Newland Formation (Schieber, 1991). Scale bar 100 ,Cm. (e) Intrabasinal shale fragment in centre of photomicrograph (sample no. 9/8/81-7). The fragment was squeezed into the space between surrounding quartz grains, an indication that it was soft when deposited. The fragment was eroded from carbonaceous swirl shale (Schieber, 1989) of the Newland Formation. Scale bar 500 ,am. (f) Closeup of intrabasinal shale fragment (sample no. 9/8/81-18). This is a piece of carbonaceous silty shale from the striped shale facies of the Newland Formation. Compare with photomicrographs of this shale type in Schieber (1986, 1989). Scale bar 100 um.

Fragments of quartzofeldspathic gneiss show granoblastic isogranular texture (Bard, 1986) and consist of quartz and feldspar crystals in the 0.1-0.5 mm size-range (Fig. 3 a). Feldspar grains with granophyric textures were also observed.

Sandstones of the Newland and Greyson formations are typically cemented by overgrowth quartz, chert and calcite. In several thin sections optically continuous overgrowth was observed on feldspar grains. Partial and even total replacement

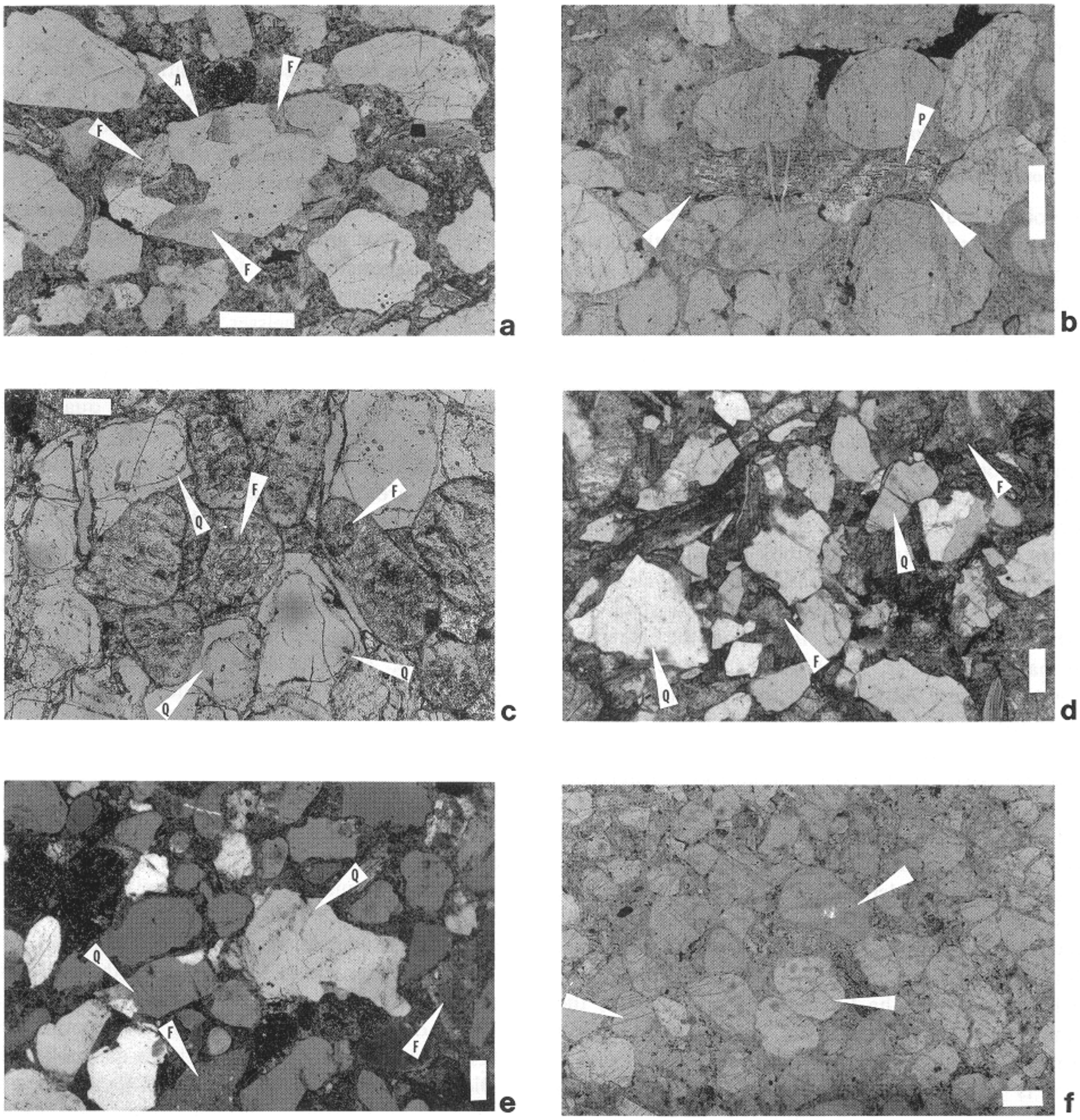

Figure 3. (a) Grain in centre of photomicrograph (arrow A) is a fragment of quartzofeldspathic gneiss (sample no. 7/5/82/6). Note granoblastic isogranular texture of quartz (light grey) and feldspar crystals (darker gray, mottled, arrows F). Scale bar 400 gym. (b) Elongate `grain' in centre (arrows) is actually the `ghost' of a quartz-replaced feldspar grain (sample no. 9/7/81-19). Traces of the feldspar original cleavage and possible ghosts of perthitic unmixing lamellae (arrow P) are still visible. Scale bar 400 um. (c) Well-rounded quartz and feldspar grains in a sandstone bed (sample no. 8/18/81-2) from the Newland Formation of the southern Little Belt mountains (Fig. 1). Quartz grains are overgrown by quartz cement in optical continuity, but are outlined by impurities along the grain boundaries (arrows Q). Feldspar grains (arrows F) have abundant inclusions due to deuteric alteration and show excellent rounding. Scale bar 100 ìm (d) Photomicrograph of typical LaHood arkose (sample no. 7/22/89-4). Shows angular quartz (arrows Q) and feldspar grains (arrows F, abundant inclusions), as well as large micas. Sorting is poor. Scale bar 200ìm (e) Photomicrograph of sandstone (sample no. 7/5/82-6) from the southern Big Belt Mountains (Fig. 1). Shows a relatively large abundance of angular quartz (arrows Q) and feldspar (arrows F) grains. Compare with Figure 3 (d). Scale bar 200 /cm (crossed polarizers). (f) Photomicrograph of sandstone in LaHood Formation (sample no. 8/20/82-3) that shows abundant well-rounded quartz grains (arrows). Scale bar 200 um.

of feldspars by calcite is a common phenomena. In several thin sections replacement of feldspar by quartz has been observed (Fig. 3 b). During point-counting all efforts were made to identify replaced feldspar grains and to record them as feldspars. Because of the considerable influence of feldspar alteration, orthoclase and plagioclase are not reported separately. Sandstones of the LaHood Formation are cemented mainly by chlorite, but also by quartz, calcite and feldspar. Partial replacement of feldspars by calcite has been observed, but is less common than in the Newland and Greyson formations.

4. Provenance information from sandstones

4.a. Source rocks

As pointed out above, sandstones of the Newland Formation are highly variable (Table 1), with detrital quartz being the most abundant and constant component. The average quartz content is 62 % but may be as small as 25 % and as large as 90 %. The feldspar content ranges from 2 % to 46 %, labile components (feldspars and rock fragments) range from 10 % to 55 %, and a wide variety of rock fragments is observed. However, close examination shows that the sedimentary rock fragments (most abundant) are identical to shale and carbonate types observed in the Newland Formation (Fig. 2d, f). The shale fragments shown in Figure 2 (d, f) are among the most common ones, and via comparison of textural features they can clearly be identified as fragments from the striped shale and carbonaceous swirl shale facies of the Newland Formation (Schieber, 1989). The carbonate fragment shown in Figure 2d is derived from calcitic stringers (so called `molar tooth structures', Smith, 1968) in fine crystalline dolostones of the Newland Formation (Schieber, 1988, 1991). These `molar tooth' fragments are found in most sandstone samples of the Newland and Greyson formation, and are easily identified because of their unique texture, consisting of clear, equidimensional calcite crystals. In a few samples, siltstone fragments that are also attributable to the Newland Formation (J. Schieber, unpub. Ph.D. diss., Univ. Oregon, Eugene, 1985) were found. Sedimentary rock fragments in the LaHood Formation have the same appearance as those in the Newland and Greyson formations, but considering that the LaHood Formation is a lateral equivalent of the latter two formations and contains equivalent shale and carbonate facies types (J. Schieber, unpub. Ph.D. diss., Univ. Oregon, Eugene, 1985; Schieber, 1989), it is most likely that these fragments were derived from within the LaHood Formation. The perfect textural match between sedimentary rock fragments in Newland sandstones and shale and carbonate units described from the Newland Formation (Schieber, 1989, 1991) leads to the obvious conclusion that all sedimentary rock fragments in the Newland Formation are of intrabasinal origin. Because in sandstone petrographical investigations of source area and tectonic setting only extrabasinal components are considered, failure to distinguish intrabasinal from extrabasinal rock fragments can lead to serious errors (Zuffa, 1985). In this study intrabasinal rock fragments were readily identified and excluded from calculation of detrital modes for QFL diagrams. If intrabasinal rock fragments are subtracted, most of the Newland samples can be classified as arkoses (5 samples) and subarkoses (21 samples), suggesting that the source terrane was dominated by granitic or gneissic basement rocks (Pettijohn, Potter & Siever, 1987). Metamorphic source rocks are also indicated by the various metamorphic rock fragments that have been observed (e.g. Fig. 3a) and by polycrystalline quartz grains with flatted quartz crystals and sutured grain boundaries (Fig. 2 a). That gneisses were a dominant source rock is also suggested by the fact that quartzofeldpathic gneiss fragments are the most common extrabasinal rock fragments. Gneisses dominate Precambrian basement rocks north as well as south of the Helena embayment (Cohenour & Kopp, 1980; Witkind, 1971 ; Keefer, 1972), and the abundance of quartzofeldspathic gneiss fragments can probably be accounted for with the observation that quartzofeldspathic gneisses are a major constituent of Precambrian basement outcrops south of the Helena embayment (Cohenour & Kopp, 1980). That granite plutons were also present in the source area is indicated by feldspar grains with granophyric and graphic textures (Fig. 2c), and by rounded inclusion of quartz in feldspar grains (Fig. 2b). The latter are described by Mehnert (1968) as `Tropfenquartz', and are attributed to anatectic granitoids.

4.b. Tectonic setting

The concept that sandstone composition reflects not only the source area but also the tectonic setting of sandstone accumulation has been expressed quite early by Krynine (1943), and has undergone considerable refinement since then (e.g. Dickinson, 1985). The QFL diagrams of Figure 4 show the distribution of detrital modes for the Newland, Greyson and LaHood formations. Applying the compositional fields for different tectonic settings as published by Dickinson (1985) indicates that all these sandstones were derived from a cratonic source as should be expected. The diagrams further show that sandstones of the LaHood Formation were primarily derived from a basement uplift, and that the Newland Formation, though dominated by sands derived from the cratonic interior, also contains an admixture of material from a basement uplift (Fig. 4). The latter circumstance is indicated by sandstones that cover the compositional range between basement uplift and cratonic interior sands in Figure 4. From the somewhat larger abundance of angular to subangular quartz grains (Table 1) in southern exposures of the Newland Formation (southern Big Belt Mountains) one might speculate that aforesaid basement uplift was located to the south. Because the LaHood Formation is a southern lateral equivalent of the Newland Formation (Schieber, 1990) this basement uplift was probably the same that gave rise to deposition of the LaHood Formation. The Greyson Formation, though petrographically similar to the

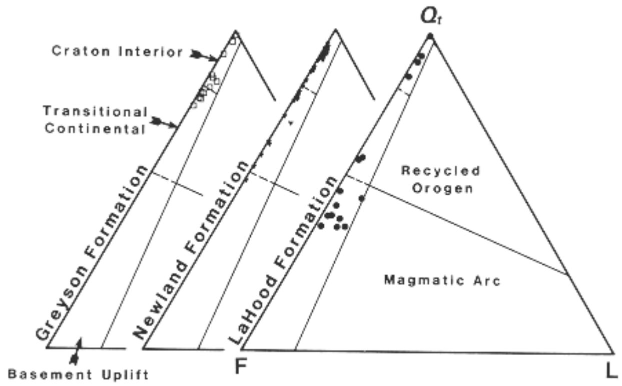

Figure 4. QFL diagrams for the sandstones in Table 1. Qt is the sum of monocrystalline quartz, polycrystalline quartz, and chert; F is the total amount of feldspars (includes altered and replaced feldspars); L is the sum of metamorphic rock fragments and quartzofeldspathic gneiss fragments of Table 1 (carbonate and terrigenous rock fragments were not counted because of intrabasinal origin). Field boundaries for the various tectonic settings are from Dickinson (1985).

Newland Formation, shows a compositional shift towards the quartz pole of the diagram. However, this shift probably does not signify a compositional change in the source rocks surrounding the Belt basin, but rather is due to erosion and recycling Newland and LaHood sediments. The latter is indicated by the abundance of Newland shale and carbonate fragments (Fig. 2d, f) and the presence of LaHood arkose fragments in Greyson sandstones. Destruction of feldspars during erosion and recycling of Newland and LaHood sediments probably was the main cause for relative quartz enrichment in Greyson sandstones.

4.c. Relief, climate, transport history

The well-rounded feldspar and quartz grains seen in most sandstone samples of the Newland Formation (Fig. 3c) indicate that the extrabasinal component of these sandstones was a texturally supermature arkose to subarkose. Folk (1980) suggests that the latter indicates tectonic quiescence, peneplanation of the hinterland, a dry climate (climatic arkose), and rounding of sand in beach or eolian environments. Even though the lack of chemically decomposed feldspars in Newland sandstones suggests a dry climate, sandstone petrography provides no clues regarding the prevailing temperature regime (hot arid v. cold arid climate). However, palaeomagnetic studies by Elston & Bressler (1980) indicate that the Belt basin was located approximately 15° north or south of the equator, and suggest in the context of observed sandstone petrology that the climate was probably hot arid to semi-arid. The high degree of rounding displayed by sandstones of the Newland Formation implies either an episode of aeolian transport (Kuenen, 1960), many cycles of transport (Pettijohn, Potter & Siever, 1987), or possibly reworking in a beach environment. Beach deposits have so far been recognized in neither the Newland Formation (J. Schieber, unpub. Ph.D. diss., Univ. Oregon, Eugene, 1985) nor its lateral equivalent, the LaHood Formation (R. L. Boyce, unpub. Ph.D. diss., Univ. Texas, Austin, 1975), and if they do exist they certainly do not form deposits of significant volume. In light of this observation and the general predominance of coastal mudflat deposits in most of the Belt Supergroup (Harrison, 1972), it seems fair to suggest that beach reworking did not contribute much to the rounding of quartz and feldspar grains in the Newland Formation. Compared to quartz, feldspars have considerably smaller chemical and mechanical stability. Therefore they are abraded faster than quartz, and their abundance is reduced greatly during successive cycles of erosion and deposition. The observation that feldspars are quite fresh and of similar average grainsize than the quartz grains suggests that the extrabasinal component of the Newland sandstones is at least in part first-cycle material. For that reason, and because it seems unlikely that beach reworking caused significant rounding of sand grains, it appears most likely that quartz and feldspar grains in Newland sandstones experienced an episode of aeolian transport.

Coarse grain-size, poor sorting, and the large abundance of angular quartz and feldspar grains in the LaHood Formation (Fig. 3d) are typical of tectonic arkoses (Folk, 1980), a characteristic sediment near basement uplifts. Feldspars in the LaHood Formation are both fresh and altered, which according to Folk (1980) would indicate a humid and warm climate, in apparent contradiction to the arid climate indicated for the Newland Formation. However, if examined in detail it becomes clear that feldspars in the LaHood Formation were most likely not altered during weathering, but rather during deuteric or hydrothermal alteration of source rocks. The following observations support that conclusion

(a) Alteration of feldspars in the LaHood Formation closely compares to alteration observed in feldspars from Precambrian basement rocks south of the Helena embayment (Casella et al.

1982; Warner, Lee-Berman & Simonds, 1982; Cordua, unpub. Ph.D. diss., Univ. Indiana, Bloomington, 1973; M. S. Duncan, unpub. Ph.D. diss., Univ. Indiana, Bloomington, 1976; P. S. Dahl, unpub. Ph.D. diss., Univ. Indiana, Bloomington, 1977; Heinrich & Rabbitt, 1960), the likely source rocks of the LaHood Formation (McMannis, 1963; R. L. Boyce, unpub. Ph.D. diss., Univ. Texas, Austin, 1975).

(b) Altered feldspars with cloudy inclusions are also common in crystalline rocks that underlie the Belt Supergroup in the Little Belt Mountains (own observations).

(c) Gneiss pebbles in LaHood Formation conglomerates show feldspar alteration of the same type as observed in detrital feldspar grains of the Newland, Greyson, and LaHood formations.

(d) Also, feldspar grains with patchy alteration that have altered and unaltered domains on the grain surface, as well as feldspar grains with altered cores and unaltered rims, indicate that alteration did not occur during weathering or diagenesis, but most likely during cooling and crystallization of igneous and metamorphic parent rocks.

5. Provenance information from shales

5.a. Source rocks

Shales are best suited for provenance studies of elastic sediments because of their relatively homogeneity, their post-depositional impermeability, and because they dominate the sedimentary mass balance. Source rock composition is commonly thought to be the dominant factor that controls their composition (Taylor & McLennan, 1985). However, secondary processes (weathering, transport, diagenesis, etc.) can have an effect on composition (Wronkiewicz & Condie, 1987; Cullers *et al.* 1987), and therefore one best relies on elements that show little mobility under the expected geological conditions. Taylor & McLennan (1985) pointed out that such elements should possess very low partition coefficients between natural waters and upper crust and short oceanic residence times. Elements that meet these criteria are, for example, Al, Ti, Th, Co and the REE (Taylor & McLennan, 1985, table 2.3).

Because of low solubility and similar behaviour during weathering, the ratio of Ti and A1 has been used as an index of provenance and as an estimator of the average bulk chemical composition of the source area (McLennan, Fryer & Young, 1979). Even though the Ti/A1 diagram as used by McLennan, Fryer & Young (1979) would suggest source rocks of granitic composition for the Newland Formation (Fig. 5), it is understood that the ratio is not sensitive to lithological provenance. It is commonly assumed that at least the bulk of the eastern Belt Supergroup was derived from the Canadian Shield (Harrison, 1972; Reynolds, 1984), and the coincidence (Fig. 5) between the Ti/AI ratio of Newland shales and the average Ti/Al ratio of the Canadian Shield (Shaw *et al.* 1967), though in itself not proving anything, is certainly consistent with this supposition.

La/Th v. Hf plots have been used to distinguish shales derived from source rocks of felsic v. mafic composition (Bhatia & Taylor, 1981; Condie & Martell, 1983). Small La/Th ratios and relatively large Hf contents of the terrigenous fraction of shales in the Newland Formation are consistent with its derivation from crust of largely granitic composition (Schieber, 1986, 1990).

5.6. Tectonic setting

It has been shown for ancient as well as for modern mudstones that their Si02 content and K20/Na20 ratio can be used to discriminate between mudstones deposited in passive margin/cratonic, active margin, and island arc tectonic settings

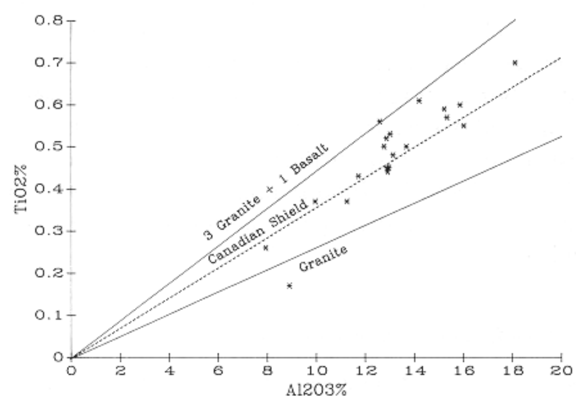

Figure 5. Ti/Al plot for shales from the upper member of the Newland Formation. The relatively small Ti/Al ratios indicate a source area dominated by rocks of granitic composition. The diagram also shows that the source area was probably of the same overall composition as the Canadian Shield. For compositional data see Table 2.

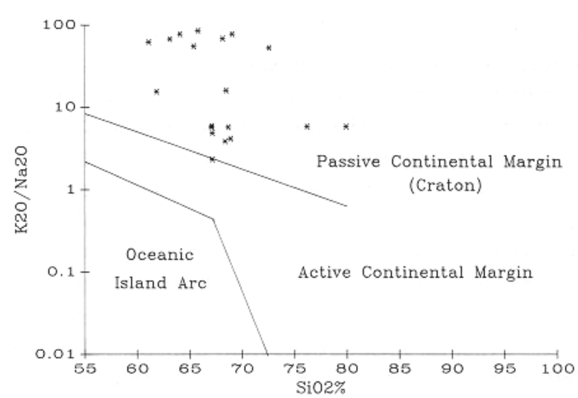

Figure 6. Si02 *v.* K20/Na20 plot of carbonate-free shales from the upper member of the Newland Formation. Field boundaries for the various tectonic settings are from Roser & Korsch (1986). For compositional data see Table 2.

Table 2. Analytical data

	Al_2O_3	CaO	Fe_2O_3	MgO	MnO	P_2O_5	K_2O	SiO_2	Na_2O	TiO_2	CIA
6/29/81-7	13.02	0.36	9.43	1.70	0.05	0.16	5.73	64.50	0.07	0.53	65.1
7/12/81-5	11.26	1.23	3.01	5.29	0.01	0.1	3.64	68.46	0.23	0.37	63.2
7/24/81-15	13.14	0.96	3.72	5.58	0.01	0.12	3.91	65.37	0.07	0.48	68.3
8/2/81-13	12.89	0.57	4.25	2.70	0.02	0.14	3.60	68.67	0.63	0.45	68.3
8/3/81-7	16.03	0.87	4.34	4.88	0.01	0.06	4.54	61.82	0.29	0.55	69.6
8/3/81-17	12.87	0.47	3.96	2.99	0.02	0.12	3.45	68.38	0.90	0.52	67.9
8/11/81-18	11.73	1.16	4.87	8.41	0.05	0.09	2.56	63.08	0.04	0.43	70.3
8/19/81-21	13.68	0.36	4.57	3.13	0.01	0.09	3.73	67.06	0.66	0.50	70.3
9/9/81-8	12.60	0.13	2.14	1.14	0.01	0.04	2.58	72.55	0.05	0.56	80.2
9/5/81-5	15.34	0.18	2.75	2.50	0.01	0.05	3.87	69.04	0.05	0.57	76.9
9/7/81-17A	18.13	0.24	5.60	2.82	0.01	0.07	5.01	61.07	0.08	0.70	75.1
9/7/81-26	12.92	0.18	4.51	5.14	0.02	0.04	2.94	68.13	0.04	0.44	78.3
9/9/81-1	15.88	0.18	4.25	2.95	0.01	0.04	4.60	65.77	0.05	0.60	74.6
8/25/81-14	15.23	0.23	4.91	3.02	0.02	0.04	3.52	67.10	0.60	0.59	74.5
8/26/81-19	12.97	0.22	4.16	6.93	0.01	0.00	2.55	67.15	0.53	0.45	76.2
8/27/81-4	12.77	0.21	3.87	5.20	0.01	0.03	2.61	68.90	0.64	0.50	75.0
8/27/81-12	9.96	0.17	2.76	4.85	0.01	0.02	2.17	76.18	0.38	0.37	75.2
8/27/81-17	7.94	0.13	2.12	4.38	0.01	0.01	1.50	79.90	0.26	0.26	77.6
8/25/81-6	8.91	1.79	1.54	3.55	0.02	0.02	2.83	77.04	0.02	0.17	58.3
9/1/82-3	14.20	0.25	6.88	2.71	0.02	0.08	3.06	67.15	1.32	0.61	70.5

All elemental abundances as percentages.

(Roser & Korsch 1986). As expected from prior knowledge on the tectonic setting of the Newland Formation, a Si02 v. K20/Na20 diagram (Fig.6) clearly shows the Newland shales to plot into the passive margin/cratonic field defined by Roser & Korsch (1986).

5.c. Climate

Various investigators have utilized the so-called `Chemical Index of Alteration' (or CIA) of Nesbitt & Young (1982) to evaluate the intensity of weathering in the source area of shales (Wronkiewicz & Condie, 1987, 1989). However, the CIA is burdened with some uncertainty because of possible post-depositional mobility of alkali and alkali earth elements. The CIA is calculated as CIA = [A1203/(A1203+CaO+NazO+K20)] x 100, where CaO denotes the calcium content of the terrigenous fraction of the sample. In carbonate-free Newland shales the CaO and Na20 content is always quite small (see Table 2). Therefore, and because K probably only experienced local redistribution (from detrital K-spar and K-mica to illite) during diagenesis (Aronson & Hower, 1976), CIA values of the Newland Formation should only show minor post-depositional modification. The average CIA value for the Newland shales (Table 2) is 71.8, suggesting moderate-to-intermediate chemical weathering (Nesbitt & Young, 1982).

6. Discussion and conclusion

In recent years it has been postulated that there was a landmass (now rifted away) to the west of the Belt basin that contributed sediment to the basin (e.g. Sears & Price, 1978 ; Winston, 1986), and one might ask if this landmass might have contributed sediment to the Newland Formation. However, if such a landmass existed, the Belt basin was probably a deep and narrow, north-trending gulf with the Newland Formation being deposited along its eastern margin (Hoffman, 1988 ; Cressman, 1989), and clastics from the western landmass were trapped within the gulf rather than being able to prograde across to the eastern margin. That there was no contribution from a western landmass is confirmed by palaeocurrent data from sandstones of the Newland Formation (Schieber, 1987). These data indicate that the sandstones were shed into the Helena embayment from the north and northeast (presumably derived from the Canadian Shield) as well as from the south (basement uplift), and are in agreement with palaeocurrent data from Newland shales (Schieber & Ellwood, 1989).

Sandstone petrography clearly indicates that the Newland Formation had both metamorphic (quartzofeldspathic gneiss, quartz muscovite schist) and intrusive rocks (micrographic and perthitic texture in feldspars) of granitic composition in the source area. Textural observations (`Tropfenquartz') indicate further that at least some of the granites owed their origin to anatexis of deeply buried metamorphic rocks. Previously, source rock information on sediments of the eastern Belt basin was only available from the study of conglomerates in the LaHood Formation. These earlier studies (Sahinen, 1950; McMannis, 1963; N. M. Gillmeister, unpub. Ph.D. diss., Univ, Harvard, 1971 ; R. L. Boyce, unpub. Ph.D. diss., Univ. Texas, Austin, 1975) led to the conclusion that the LaHood source area was dominantly a gneissic terrane with local occurrences of amphibolites, schists, and marbles. These conclusions agree well with the source rock assessment for sandstones in the Newland Formation, which is of course only as it should be, considering the commonly assumed lateral equivalency between the Newland and LaHood formations (McMannis, 1963 ; R. L. Boyce, unpub. Ph.D. diss., Univ. Texas, Austin, 1975; Schieber, 1990). Marble fragments, if misidentified as intrabasinal carbonate fragments, can potentially cause errors in QFL plots. However, in the studied rocks they have a unique strained and coarse crystalline appearance, are exceedingly rate (only one grain was counted in 55 samples), and are therefore of no importance to QFL plots. Shale geochemistry of the Newland Formation also indicates a source area of overall granitic composition. A Ti/Al ratio that is comparable to the Canadian Shield and general considerations about the composition of the upper continental crust (Taylor & McLennan, 1985) suggest a source area dominated by granitoid gneisses. However, in contrast to sandstone petrology, shale geochemistry does not provide a *de facto* distinction between metamorphic and igneous source rocks.

Work on the LaHood Formation also led to the recognition that its sediments were derived from uplifted basement rocks along the southern, fault-controlled, margin of the Helena embayment (Sahinen, 1950; McMannis, 1963; N. M. Gillmeister, unpub. Ph.D. diss., Univ, Harvard, 1971), a conclusion that is confirmed by detrital modes of LaHood sandstones in a QFL diagram (Fig. 4). In contrast, detrital modes of most LaHood sandstones examined by R. L. Boyce (unpub. Ph.D. diss., Univ. Texas, Austin, 1975) plot in the magmatic arc and recycled orogen field of Dickinson (1985). Boyce had used the traditional point-counting method (rather than Gazzi-Dickinson), and after comparing my thin sections with rock descriptions by Boyce I concluded that he had misidentified badly altered feldspars as metamorphic rock fragments, thus causing a considerable shift of datapoints towards the rock fragment pole of the diagram. Because of the lateral equivalency of the LaHood and Newland formations one should expect to find abundant angular quartz and feldspar grains at least in those locations of the Newland Formation that are close to the LaHood outcrop belt (Newland Formation of the southern Big Belt Mountains, Fig. 1). However, even though a few samples do indeed show a large abundance of angular quartz and feldspar (Fig. 3 e), and notwithstanding that palaeoflow indicators show a derivation from the south (LaHood Formation), samples from the southern Big Belt Mountains (Fig. 1) are strongly dominated by well rounded quartz and feldspar grains (85 % well rounded grains on average), indicative of derivation largely from a peneplaned cratonic interior source. Two possible scenarios come to mind to reconcile the obvious discrepancy between observation and expectation: (1) sediment from the cratonic surface bypassed the southern basement uplift through (possibly fault-related) valleys; (2) longshore drift moved well-rounded sediments from other parts of the embayment to the shorelines of the LaHood Formation, and storms moved this material deeper into the basin. Studies of Precambrian tectonics by Schmidt & Garihan (1986) lend some support to scenario (1). However, the presence of beds of well-rounded sandstone in nearshore facies of the LaHood Formation (Fig. 2f) and the absence thereof in proximal (close to fault zone, Fig. 1) LaHood sediments (personal observations), as well as the observation that Newland sandstones of the southern Big Belt Mountains were carried into the basin from the south by storm-induced currents (Schieber, 1987) suggests to me that scenario (2) is more likely.

Where a sandstone plots on a QFL diagram does not only depend on the tectonic setting of the source area (Dickinson, 1985), but also on the prevailing climate (Basu, 1985; Cullers *et al.* 1987; Cullers, 1988). In particular sandstones derived from low-to moderate relief source areas under humid climatic conditions may be depleted in feldspars and rock fragments and enriched in quartz (Potter, 1978; Franzinelli & Potter, 1983; Mack, 1984; Basu, 1985). Feldspar preservation and palaeomagnetic studies (see above) indicate a hot and arid climate during Newland deposition. Under these conditions, even though a low-relief source area is likely for most of the Newland Formation, the effect of climate on the modal composition of sandstones is minimal and will most likely not cause errors in the interpretation of tectonic setting from QFL diagrams. Sandstone petrology and shale geochemistry lead to the same conclusion with regard to the tectonic setting of the Newland Formation, namely that it was deposited in a passive margin or intracratonic setting. Because of recent palaeogeographical reconstructions of Mid-Proterozoic continents and the Belt basin (Stewart, 1976; Sears & Price, 1978; Piper, 1982; Winston, 1986; Cressman, 1989), an intracratonic setting appears most likely.

A considerable discrepancy between conclusions reached from sandstone petrography v. shale geochemistry exists with regard to the climate at the time of Belt deposition. Whereas sandstone petrography suggests a peneplaned craton, dry climate, and very limited or absent chemical weathering (climatic arkose of Folk, 1980), shale geochemistry (CIA) indicates a moderate-to-intermediate degree of chemical weathering. Actually, the abundant presence of shale itself poses another conundrum, because it is commonly assumed that soil (and clay) formation is very limited under conditions that favour climatic arkoses (Folk, 1980). Aside from indicating that the CIA as used to determine weathering intensity in the source areas of shales is in need of reassessment and refinement, the conflicting climate signals between sandstones and shales of the Newland Formation may also be inherent in the way sandstones and shales are produced. For example, sediment transport can, via hydraulic sorting, lead to compositional fractionation of sediment-size fractions (Nesbitt, 1979 ; Reimer, 1985 ; Cullers et al. 1987). As pointed out above, unaltered feldspars (insignificant chemical weathering) and well rounded quartz and feldspar grains (aeolian reworking) in conjunction with palaeomagnetic data (low latitude) suggest that the Newland Formation was deposited in a hot arid to semi-arid climate. In such a climatic setting unaltered feldspars would become concentrated in sandy deposits of braided streams and may also undergo aeolian reworking, whereas fine detritus and clay (from feldspar weathering) would be carried to the basin as suspended load, thus leading to a separation between intensely (clay fraction) and incompletely weathered (sand fraction) material. Therefore intensities of chemical weathering indicated by shales will tend to be higher than those indicated by sandstones. It appears therefore that for realistic estimates of source area weathering conditions the data from shales and sandstones should be considered in conjunction. One approach to better understand the complexities of this problem would be a study of modern sandstones in conjunction with a study of the connection between composition and climate for modern mudstones.

7. Summary

Conclusions that can be drawn from this study address on one hand particular questions of the provenance of the Newland Formation, and on the other hand the degree of resolution and congruence between conclusions based on sandstone petrography v. shale geochemistry.

Conclusions that pertain to Newland Formation provenance are

(1) The Newland Formation was derived from a source area that was dominated by granitoid gneisses but also contained an appreciable proportion of granites.

(2) The Newland Formation accumulated in an intracratonic tectonic setting and received sediment from the Canadian Shield to the northeast as well as from a basement uplift along the southern basin margin.

(3) Most of the hinterland had undergone peneplanation and, as geochemical data of shales indicate, the climate was probably not quite as arid as sandstone petrology would suggest. A semi-arid climate with moderate amounts of chemical weathering is assumed.

Conclusions that pertain to provenance information obtainable from shales and sandstones are:

(1) With respect to source rock identification and tectonic setting, both sandstone petrology and shale geochemistry lead to the same general result. None the less, information obtained from sandstone petrology clearly allows a more detailed assessment.

(2) The estimate of weathering intensity that is obtained from sandstone petrology is notably different from that derived from shale geochemistry. The reason for this discrepancy deserves further investigation.

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References

ARONSON, J. L. & HOWER, J. (1976). Mechanism of burial metamorphism of argillaceous sediments. 2. Radiogenic argon evidence. *Geological Society of America Bulletin* 87, 738-44.

BARD, J. P. 1986. *Microtextures of Igneous and Metamorphic Rocks.* Dordrecht: D. Reidel, 264 *pp.*

BAsu, A. 1985. Influence of climate and relief on compositions of sandstones released at source areas. In *Provenance of Arenites* (ed. G. G. Zuffa), pp. 1-18. Dordrecht: D. Reidel.

BASU, A., YOUNG, S. W., SUTTNER, L. J., JAMES, W. C. & MACK, G. H. 1975. Re-evaluation of the use of undulatory extinction and polycrystallinity in detrital quartz for provenance interpretation. *Journal of Sedimentary Petrology 45*, 873-82.

BHATIA, M. R., TAYLOR, S. R. 1981. Trace element geochemistry and sedimentary provinces: study from the Tasman geosyncline, Australia. *Chemical Geology 33,* 115-25.

B16RLYKKE, K. 1974. Geochemical and mineralogical influence of Ordovician island arcs on epicontinental elastic sedimentation. A study of Lower Paleozoic sedimentation in the Oslo region, Norway. *Sedimentology 21,* 251-72.

CASELLA, C. J., LEVAY, J., EBLE, E., HIRST, B., HUFFMAN, K., LAHTI, V. & METZGER, R. 1982. Precambrian geology of the southwestern Beartooth Mountains, Yellowstone National Park, Montana and Wyoming. *Montana Bureau of Mines and Geology Special Publication 84,* 1-24.

COHENOUR, R. E. & Kopp, R. S. 1980. *Regional Investigation for Occurrence of Radioactive Quartz-pebble Conglomerates in the Precambrian of' Southwestern Montana.* National Uranium Resource Evaluation Final Report, United States Department of Energy, 582 pp.

CONDIE, K. C. & MARTELL, C. 1983. Early Proterozoic metasediments from north-central Colorado. *Geological Society of America Bulletin 94,* 1215-24.

CRESSMAN, E. R. 1989. *Reconnaissance Stratigraphy of the Prichard Formation (Middle Proterozoic) and the Early Development of the Belt Basin, Washington, Idaho, and Montana.* United States Geological Survey Professional Paper, no. 1490, 80 *pp.*

CULLERs, R. L. 1988. Mineralogical changes of soil and stream sediment formed by intense weathering of the Danburg granite, Georgia, *U.S.A. Lithos 21,* 301-14.

CULLERS, R. L., BARRETT, T., CARLSON, A. & ROBINSON, B. 1987. REE and mineralogic changes in Holocene soil and stream sediment: a case study in the Wet Mountains, Colorado, U.S.A. *Chemical Geology 63,* 275-97.

DICKINSON, W. R. 1985. Interpreting provenance relations from detrital modes of sandstones. In *Provenance of Arenites* (ed. G. G. Zuffa), pp. 333-61. Dordrecht: D. Reidel.

ELSTON, D. P. & BRESSLER, S. L. 1980. Paleomagnetic poles and polarity zonation from the Middle Proterozoic Belt Supergroup, Montana and Idaho. *Journal of Geophysical Research 85,* 339-55.

ENGLUND, J.-O. & JORGENSEN, P. 1973. A chemical classification system for argillaceous sediments and factors affecting their composition. *Geologiska Foreningens i Stockholm Fohandlingar 95,* 87-97.

FOLK, R. L. 1968. Bimodal supermature sandstones: products of the desert floor. *XXlllrd International Geological Congress Proceedings 8,* 9-32.

FOLK, R. L. 1980. *Petrology of Sedimentary Rocks.* Austin, Texas: Hemphill Publication Company, 182 pp.

FRANZINELLI, E. & POTTER, P. E. 1983. Petrology, chemistry and texture of modern river sands, Amazon River system. *Journal of Geology 91,* 23-39.

HARRISON, J. E. 1972. Precambrian Belt basin of northwestern United States: its geometry, sedimentation, and copper occurrences. *Geological Society of America Bulletin 83,* 1215-40.

HEINRICH, E. W. & RABBITT, J. C. 1960. Pre-Beltian geology of the Cherry Creek and Ruby Mountain areas, southwestern Montana. 1. Geology of the Cherry Creek area, Gravelly Range, Madison County, Montana. *Montana Bureau of Mines and Geology Memoir 38,* 1-14.

HOFFMAN, P. F. 1988. Belt basin: landlocked remnant oceanic basin? (Analogous to the South Caspian and Black Seas.) *Geological Society of America Abstracts with Programs 20,* A 50.

INGERSOLL, R. V., BULLARD, T. F., FORD, R. L., GRIMM, J. P., PICKLE, J. D. & SARES, S. W. 1984. The effect of grain size on detrital modes: a test of the GazziDickinson point-counting method. *Journal of Sedimentary Petrology 54,* 103-16.

KEEFER, W. R. 1972. *Geological Map of the West Half of the Neihart Quadrangle, Montana.* United States Geological Survey Miscellaneous Geological Investigations, Map 1-726.

KRYNINE, P. D. 1943. *Diastrophism and the Evolution of Sedimentary Rocks.* Pennsylvania Mining Industry Technical Paper no. 84-A, 21 pp.

KUENEN, P. H. 1960. Experimental abrasion. 4. Eolian action. *Journal of Geology 68,* 427-49.

MACK, G. H. 1984. Exceptions to the relationship between plate tectonics and sandstone composition. *Journal of Sedimentary Petrology 54,* 212-20.

McBRIDE, E. F. 1985. Diagenetic processes that affect provenance determinations in sandstone. In *Provenance of Arenites* (ed. G. G. Zuffa), pp. 95-113. Dordrecht: D. Reidel.

MCBRIDE, E. F. 1987. Diagenesis of the Maxon Sandstone (Early Cretaceous), Marathon region, Texas. *Journal of Sedimentary Petrology 57,* 98-107.

MCLENNAN, S. M., FRYER, B. J. & YOUNG, G. M. 1979. The geochemistry of the carbonate-rich Espanola Formation (Huronian) with emphasis on the rare earth elements. *Canadian Journal of Earth Sciences 16,* 230-9.

MCMANNIS, W. J. 1963. LaHood Formation-a coarse facies of the Belt Series in southwestern Montana. *Geological Society of America Bulletin 74,* 407-36.

MEHNERT, K. R. 1968. *Migmatites and the Origin of Granitic Rocks.* Amsterdam: Elsevier, 405 pp.

NESBITT, R. W. 1979. Mobility and fractionation of REE during weathering of a granodiorite. *Nature 279,* 206-10.

NESBITT, R. W. & YOUNG, G. M. 1982. Early Proterozoic climates and plate motions inferred from major element chemistry of lutites. *Nature 299,* 715-17.

PETTIJOHN, F. J., POTTER, P. E. & SIEVER, R. 1987. *Sand and Sandstone.* New York: Springer Verlag, 553 pp.

PIPER, J. D. A. 1982. The Precambrian paleomagnetic record: the case for the Proterozoic supercontinent. *Earth and Planetary Science Letters 59,* 61-89.

PITTMAN, E. D. 1970. Plagioclase as an indicator of provenance in sedimentary rocks. *Journal of Sedimentary Petrology 40,* 591-8.

POTTER, P. E. 1978. Petrology and chemistry of modern big river sands. *Journal of Geology 86,* 423-49.

POWERS, M. C. 1953. A new roundness scale for sedimentary particles. *Journal of Sedimentary Petrology 23,* 117-19.

REIMER, T. O. 1985. Rare earth elements and the suitability of shales as indicators for the composition of the Archean continent crust. *Neues Jahrbuch fur Mineralogie Abhandlungen 152,* 211-23.

REYNOLDS, M. E. 1984. Tectonic setting and development of the Belt basin, northwestern United States. *Montana Bureau of Mines and Geology Special Publication 90,* 44-6.

ROSER, B. P. & KORSCH, R. J. 1986. Determination of tectonic setting of sandstone-mudstone suites using Si02 content and K20/Na20 ratio. *Journal of Geology* 94, 635-50.

SAHINEN, U. M. 1950. *Geology and Ore Deposits of the Highland Mountains, Southwest Montana.* Montana Bureau of Mines and Geology, Memoir no. 32, 63 pp.

SCHIEBER, J. 1986. Stratigraphic control of rare-earth pattern types in Mid-Proterozoic sediments of the Belt Supergroup, Montana, U.S.A.: implications for basin analysis. *Chemical Geology* 54, 135-48.

SCHIEBER, J. 1987. Storm-dominated epicontinental clastic sedimentation in the Mid-Proterozoic Newland Formation, Montana, U.S.A. *Neues jahrbuch fur Geologie and Paldontologie Monatshefte, pp.* 417-39.

SCHIEBER, J. 1988. Redistribution of rare-earth elements during diagenesis of carbonate rocks from the MidProterozoic Newland Formation, Montana, U.S.A. *Chemical Geology* 69, 111-26.

SCHIEBER, J. 1989. Facies and origin of shales from the MidProterozoic Newland Formation, Belt Basin, Montana, U.S.A. *Sedimentology* 36, 203-19.

SCHIEBER, J. 1990. Distribution of REE in the eastern Belt Supergroup (Montana, U.S.A.): implications for stratigraphic correlations and basin evolution. *Chemical Geology 81,* 83-98.

SCHIEBER, J. 1991. Facies and deposition of a mixed terrigenouscarbonate suite in a Mid-Proterozoic Epicratonic sea: the Newland Formation, Belt Supergroup, Montana, U.S.A. *Neues Jahrbuch fur Geologie and Palaontologie Monatshefte* (in press).

SCHIEBER, J. & ELLWOOD, B. B. 1989. Application of anisotropy of magnetic susceptibility (AMS) to paleocurrent studies in shale sequences. EOS 70, 1066.

SCHMIDT, C. J. & GARIHAN, J. M. 1986. Middle Proterozoic and Laramide tectonic activity along the southern margin of the Belt basin. *Montana Bureau of Mines and Geology Special Publication 94,* 217-35.

SEARS, J. W. & PRICE, R. A. 1978. The Siberian connection a case for Precambrian separation of the North American and Siberian cratons. *Geology 6,* 267-70.

SHAW, D. M., REILLY, G. A., MUYSSON, J. R., PATTENDEN, G. E. & CAMPBELL, F. E. 1967. An estimate of the chemical composition of the Canadian Precambrian shield. *Canadian Journal of Earth Sciences 4,* 829-53.

SMITH, A. G. 1968. The origin and deformation of some `Molar Tooth Structures' in the Precambrian Belt Purcell Supergroup. *Journal of Geology 76,* 426-43.

STEWART, J. H. 1976. Late Precambrian evolution of North America: plate tectonics implication. *Geology 4,* 11-15.

TAYLOR, S. R. & MCLENNAN, S. M. 1985. *The Continental Crust: Its Composition and Evolution.* Oxford: Blackwell, 312 pp.

WARNER, J. L., LEE-BERMAN, R. & SIMONDS, C. H. 1982. Field and petrologic relations of some Archean rocks near Long Lake, eastern Beartooth Mountains, Montana and Wyoming. *Montana Bureau of Mines and Geology Special Publication 84,* 56-68.

WINSTON, D. 1986. Sedimentology of the Ravalli Group, Middle Belt Carbonate and Missoula Group, Middle Proterozoic Belt Supergroup, Montana, Idaho and Washington. *Montana Bureau of Mines and Geology Special Publication 94,* 85-124.

WITKIND, 1. J. 1971. *Geologic Map of the Barker Quadrangle Judith Basin and Cascade Counties, Montana.* United States Geological Survey, Map GQ-898.

WRONKIEWICZ, D. J. & CONDIE, K. C. 1987. Geochemistry of Archean shales from the Witwatersrand Supergroup, South Africa: source-area weathering and provenance. *Geochimiea et Cosmochimica Acta 51,* 2401-16.

WRONKIEWICZ, D. J. & CONDIE, K. C. 1989. Geochemistry and provenance of sediments from the Pongola Supergroup, South Africa: evidence for a 3.0-Ga-old craton. *Geochimica et Cosmochimica Acta 53,* 1537-49.

ZUFFA, G. G. 1985. Optical analyses of arenites: influence of methodology on compositional results. In *Provenance of Arenites* (ed. G. G. Zuffa), pp. 165-89. Dordrecht: D. Reidel.