

# Chemostratigraphy of predominantly siliciclastic Neoproterozoic successions: a case study of the Pocatello Formation and Lower Brigham Group, Idaho, USA

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**Abstract** – Isotopic chemostratigraphy has proven successful in the correlation of carbonate-rich Neoproterozoic successions. In successions dominated by siliciclastic rocks, chemostratigraphy can be problematic, but if thin carbonates punctuate siliciclastic strata, useful isotopic data may be obtained. The upper Pocatello Formation and lower Brigham Group of southeastern Idaho provide an opportunity to assess the potential and limitations of isotopic chemostratigraphy in overwhelmingly siliciclastic successions. The 5000 m thick succession consists predominantly of siliciclastic lithologies, with only three intervals that contain thin intercalated carbonates. Its depositional age is only broadly constrained by existing biostratigraphic, sequence stratigraphic and geochronometric data. The lowermost carbonates include a cap dolomite atop diamictites and volcanic rocks of the Pocatello Formation. The  $\delta^{13}\text{C}$  values of these carbonates are distinctly negative ( $-5$  to  $-3$ ), similar to carbonates that overlie Neoproterozoic glaciogenic rocks worldwide. Stratigraphically higher carbonates record a major positive  $\delta^{13}\text{C}$  excursion to values as high as  $+8.8$  within the carbonate member of the Caddy Canyon Quartzite. The magnitude of this excursion is consistent with post-Sturtian secular variation recorded elsewhere in the North American Cordillera, Australia, Svalbard, Brazil and Namibia, and exceeds the magnitude of any post-Varanger  $\delta^{13}\text{C}$  excursion documented to date. In most samples, Sr-isotopic abundances have been altered by diagenesis and greenschist facies metamorphism, but a least-altered value of approximately 0.7076 supports a post-Sturtian and pre-Marinoan/Varanger age for upper Pocatello and lower Brigham rocks that lie above the Pocatello diamictite. Thus, even though available chemostratigraphic data are limited, they corroborate correlations of Pocatello Formation diamictites and overlying units with Sturtian glaciogenic rocks and immediately post-Sturtian successions in western North America and elsewhere.

## 1. Introduction

The stratigraphic investigation of Neoproterozoic (545–1000 Ma) successions in western North America has been energized in recent years by three significant lines of research. Firstly, the recognition of sequence packages has permitted hypotheses of regional correlation despite poor biostratigraphic control (Christie-Blick & Levy, 1989*b*; Levy & Christie-Blick, 1991*a*). Secondly, Aitken's (1991) discovery of two stratigraphically distinct glacial horizons in the Mackenzie Mountains of northwest Canada demonstrates that both Sturtian (Rapitan) and Marinoan/Varanger (Ice Brook) ice ages are recorded in western North America. Finally, the SWEAT hypothesis, proposed by Moores (1991) and elaborated by Dalziel (1991), Hoffman (1991), Young (1992) and others, has suggested comparisons of Australian, Antarctic and North American successions that are geographically widely separated at present, but which may have been closely positioned during the late Neoproterozoic Era. Powerful hypotheses can be generated following each of these lines of research, and in all cases testing

requires the development of an independent chronostratigraphic framework.

The Pocatello Formation and Brigham Group of southeastern Idaho and northern Utah (Fig. 1) comprise a thick succession deposited during and after Neoproterozoic rifting and subsidence (Fig. 2; Link *et al.* 1985, 1987, 1993; Christie-Blick, Grotzinger & Von der Borch, 1988; Christie-Blick & Levy, 1989*a*; Levy & Christie-Blick, 1991*b*; Link & Smith, 1992). Although the succession is broadly constrained to be Neoproterozoic to lowest Cambrian in age, the age of these rocks is known only from sparse radiometric dates from sequences thought to be correlative, and from interpolation between limited lithostratigraphic and palaeontological tie points. In this paper we attempt to constrain further the age of this sequence through the use of established patterns of temporal variation in the C- and Sr-isotopic composition of Neoproterozoic seawater.

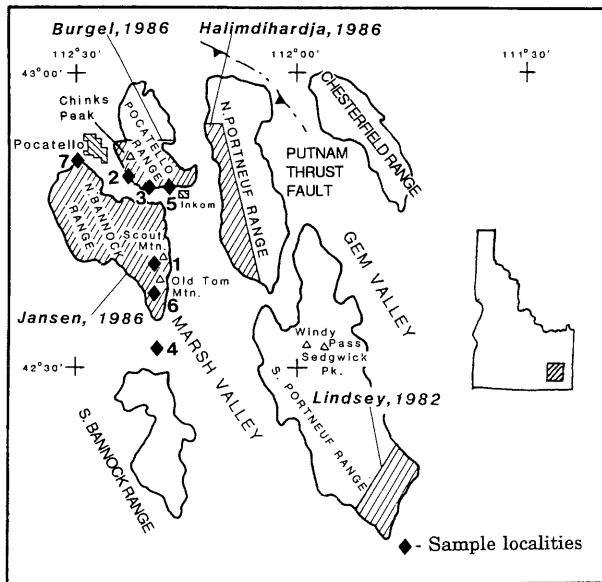


Figure 1. Location map with names of ranges and localities, after Link *et al.* 1987. Localities: 1. Scout Mountain Road, elevation 8480', just east of knob 8634. NE 1/4 NW 1/4 Section 3, T9S, R35E, 42° 40' 10" N, 112° 20' 30" W, Scout Mountain 7 1/2 minute quadrangle, Idaho. 2. North of Portneuf Narrows, elevation 5600', on east side of ridge north of Portneuf Narrows, at top of slope above the large mine. SE 1/4 SW 1/4 Section 15, T7S, R35E, 42° 48' 30" N, 112° 21' 15" W; Inkom 7 1/2 minute quadrangle; Link (1983, 1987), Link & LeFebre (1983) and Trimble (1976). 3. Portneuf Narrows east of locality 2, elevation 4760, SE 1/4 NE 1/4 Section 22, T7S, R35E, 42° 47' 45" N, 112° 20' 30" W; Inkom 7 1/2 minute quadrangle. 4. Cedar Mountains, north of Virginia Road, elevation 5320', NW 1/4 NW 1/4 Section 36, T10S, R35E, 42° 31' 15" N, 112° 18' 30" W; Hawkins 7 1/2 minute quadrangle; Platt (1985). 5. North of rest area on Interstate-15 (northbound), elevation 4920', at end of small ridge. NW 1/4 NE 1/4 Section 19, T7S, R36E, 42° 48' 30" N, 112° 17' 15" W; Inkom 7 1/2 quadrangle; Kellogg (1992) and Trimble (1976). 6. Old Tom Mountain, north of Garden City Gap, elevation 7520', NW 1/4 NW 1/4 Section 36, T9S, R35E, 112° 18' 30" W, 42° 36' 15" N; Hawkins 7 1/2 minute quadrangle; Platt (1985). 7. Saddle west of peak 5714, SE 1/4, NE 1/4, Section 21 T7S, R34E, 42° 52' N, 112° 28' W; Pocatello South 7 1/2 minute quadrangle; Trimble (1976). Hatching indicates areas of Neoproterozoic rocks recently studied (unpublished Master's theses, Pocatello, Idaho State University). Samples are indexed to localities in Table 1.

## 2. Lithostratigraphy and age constraints

In southeastern Idaho and northern Utah, Neoproterozoic deposition is represented by thick, predominantly siliciclastic sediments (Fig. 2). The sequence generally contains a diamictite and volcanic succession (the Pocatello Formation) overlain by a terrigenous detrital succession (the Brigham Group). Recent studies of these rocks include those of Crittenden, Christie-Blick & Link (1983), Link *et al.* (1985, 1987, 1993), Christie-Blick, Grotzinger & Von der Borch (1988), Christie-Blick & Levy (1989*a*), Levy & Christie-Blick (1991*a, b*), Link & Smith (1992), and

Levy, Christie-Blick & Link (in press). This lower greenschist grade succession is part of the Paris–Putnam–Willard thrust sheet of the Idaho–Wyoming thrust belt (Link *et al.* 1985; Kellogg, 1990, 1992). The carbonate rocks sampled here all come from localities in the Bannock Range (Fig. 1), with the exception of one sample from Hardscrabble Canyon in Utah (see section 3).

### 2.a. Diamictite and volcanic succession (Pocatello Formation)

The diamictite and volcanic succession of the Pocatello Formation and the correlative formation of Perry Canyon in Utah (Anderson, 1928; Ludlum, 1942; Trimble, 1976; Link, 1981, 1983, 1986, 1987; Crittenden, Christie-Blick & Link, 1983; Link & LeFebre, 1983; Christie-Blick, 1985) is extensively exposed in the allochthonous Putnam, Paris and Willard thrust sheets of Utah and Idaho. The Bannock Volcanic Member contains as much as 1000 m of basaltic volcanoclastic rocks and poorly sorted detrital strata, including greywacke and diamictite. Stratigraphic and geochemical data suggest that this succession was deposited within an extensional tectonic setting, probably related to the early stages of continental rifting (Harper & Link, 1986). Sedimentary facies associations suggest marine deposition in basins at least tens of kilometres across, and influenced by contemporaneous volcanism and faulting (Link, 1983, 1986; Christie-Blick, 1985). However, unequivocal Neoproterozoic faults have not been identified.

The Scout Mountain Member of the Pocatello Formation contains diamictite representing two distinct pulses of glaciation (Crittenden, Christie-Blick & Link, 1983). The lower diamictite has been interpreted to be a subaqueous mass flow deposit, whereas the upper diamictite is interpreted as a subaqueous flow tillite, deposited close to the glacial margin (Link, 1983). The upper diamictite is overlain by a distinctive pink dolomite caprock (pink dolomite unit, Crittenden, Christie-Blick & Link, 1983; Link & LeFebre, 1983) similar to cap limestones and dolostones above both Varanger and Sturtian glacial diamictites in many other parts of the world. An erosional unconformity separates the pink dolomite from the overlying sandstone. At its base, the sandstone unit contains a flat pebble conglomerate of large (25–50 cm wide by 2–5 cm thick) dolomite clasts, supported by a coarse red sandy matrix. This unit is overlain by the upper sandstone and limestone unit, which fines upward from coarse sandstone to argillite and limestone. A distinctive silty limestone or marble marks the top of the Scout Mountain Member. The thick (~700 m) upper member of the Pocatello Formation contains laminated slaty to phyllitic argillite, with brown quartzite interbeds increasing in frequency towards the top of the member. The base of

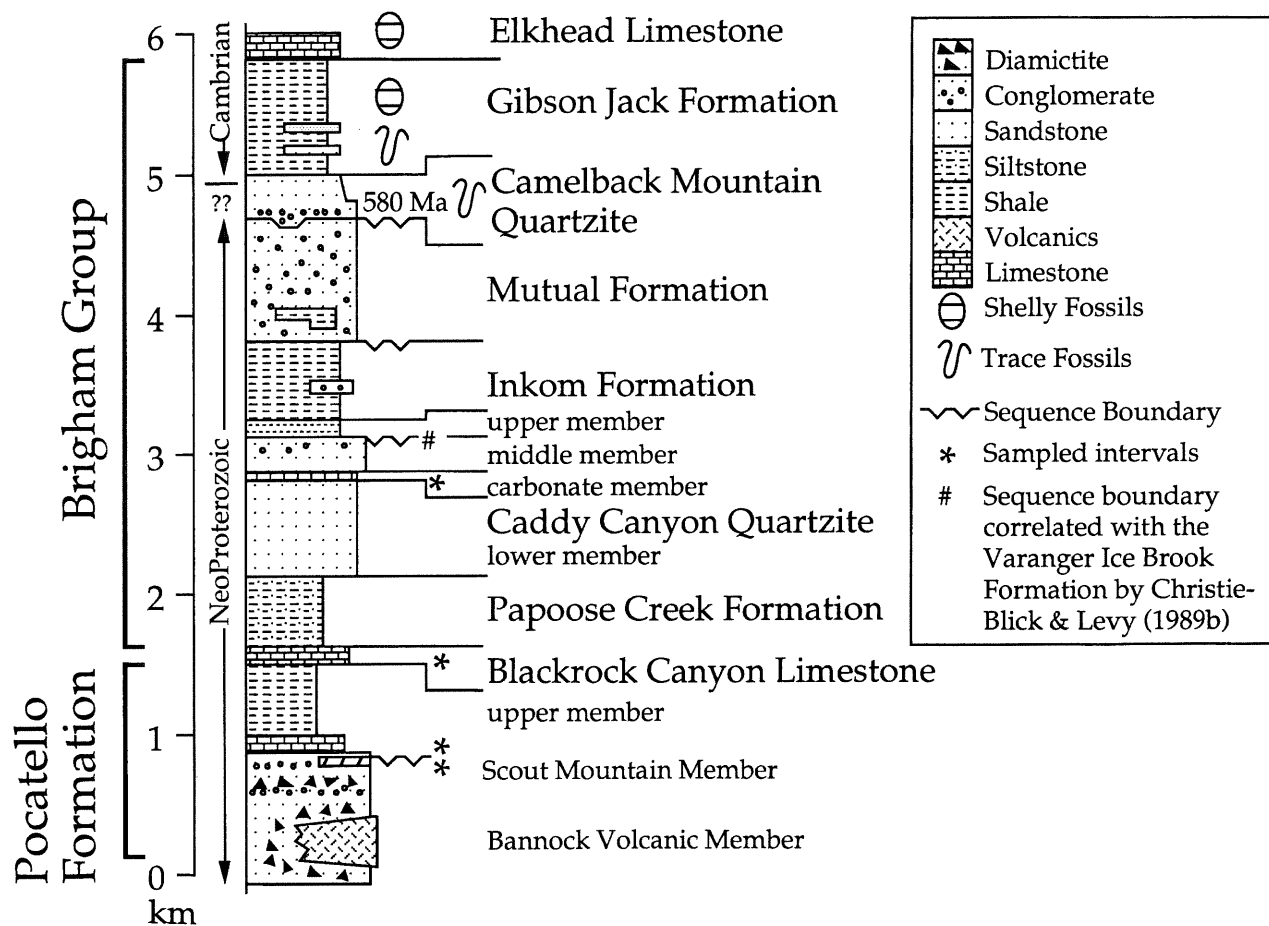


Figure 2. Generalized Neoproterozoic stratigraphy of the Pocatello Formation and Brigham Group, northern Bannock and Portneuf ranges, Idaho. 580 Ma – age from a thin tuff bed in the correlative Brown’s Hole Formation, Utah (recalculated by N. Christie-Blick from Crittenden & Wallace, 1973; no error given).

the intraclast conglomerate and sandstone above the pink dolomite unit is interpreted as an unconformity, sequence boundary, and transgressive surface, and the strata above the sequence boundary are thought to have accumulated in a shallow marine environment during the post-glacial transgression (Link, 1983; Link & LeFebre, 1983; Link & Smith, 1992). Samples were taken from each carbonate horizon at several localities (Fig. 1) and stratigraphic levels (Fig. 2).

**2.b. Blackrock Canyon Limestone**

The Blackrock Canyon Limestone is a locally developed unit (30+ m thick) that conformably overlies the upper member of the Pocatello Formation (Trimble, 1976). The formation contains several oolitic and flat-laminated limestone units separated by thin terrigenous beds. The contact between the Blackrock Canyon Limestone and the overlying Papoose Creek Formation of the Brigham Group is gradational. Samples were taken at approximately 10 m intervals through the unit at locality 3 (Fig. 1).

**2.c. Terrigenous detrital succession (Brigham Group)**

The Brigham Group of northern Utah and south-

eastern Idaho is represented by 2000 to 4000 m of quartzose sandstone, with subordinate conglomerate, mudstone, limestone and volcanic rock (Fig. 2). These rocks traditionally are called ‘quartzite’, and have generally been subjected to low-grade metamorphism (lower greenschist facies or below). Sediments of the Brigham Group are mineralogically and texturally mature, and were deposited in braided fluvial and shallow-marine environments, with occasional off-shore deposition. The succession is interpreted as having accumulated either on a passive-margin (Stewart, 1972; Stewart & Suczec, 1977; Link *et al.* 1987) or in an intracratonic basin, with a passive margin represented only by strata in the uppermost part of the sequence (Devlin & Bond, 1988; Christie-Blick & Levy, 1989b).

The Brigham Group in the Bannock Range of southeastern Idaho contains six formations (Link *et al.* 1985). The Papoose Creek Formation at the base of the Brigham Group consists predominantly of thin-bedded green, brown and grey fine-grained sandstone and siltstone, with numerous intervals containing subaqueous shrinkage cracks. The overlying Caddy Canyon Quartzite contains quartz arenite, with minor siltstone, conglomerate, pebbly sandstone, limestone

and dolomite; it is divided into four informal members. The lower member contains 500 to 600 m of planar and trough stratified quartz sandstone. The sampled carbonate member comprises approximately 10–15 m of grey limestone and brown dolomite. The middle and upper members contain poorly sorted, trough cross-bedded quartzite and conglomerate. The remaining succession within the Brigham Group (Inkom Formation, Mutual Formation and Camelback Mountain Quartzite) represents continued deposition along the episodically subsiding western margin of North America. The overlying rocks are poorly exposed, and are included in the Cambrian Gibson Jack Formation.

Recent workers have used sequence stratigraphy to elucidate the depositional history of the Brigham Group and to facilitate regional and inter-regional correlation (Christie-Blick & Levy, 1985, 1989*a*; Link *et al.* 1987; Levy & Christie-Blick, 1991*a*; Levy, Christie-Blick & Link, in press). Several unconformities of regional extent have been recognized, and at least five sequence boundaries have been inferred to date. The oldest contains the glacial deposits and the youngest represents the lower part of the Cambrian–Ordovician Sauk I sequence of the craton (Sloss, 1988). Tentative correlations have been made with other Neoproterozoic sequences along the western edge of the craton (Christie-Blick & Levy, 1989*b*; Levy, Christie-Blick & Link, in press).

#### 2.d. Metamorphic history

The Neoproterozoic through Cambrian rocks of the Pocatello Formation and Brigham Group have undergone several phases of deformation (Burgel, Rodgers & Link, 1987). Compression and regional metamorphism of the Sevier orogeny during Cretaceous time was followed by faulting and uplift associated with Neogene Basin and Range extension, and thermal elevation and subsidence caused by the passing of the Snake River Plain hot spot (Pierce & Morgan, 1992).

In the Portneuf Narrows area (locality 2), pelitic rocks of the Scout Mountain Member contain the lower greenschist facies mineral assemblage chlorite–epidote–albite, which forms a cleavage imparted during Mesozoic thrusting (Link, 1983; Snoke & Miller, 1988). Subjacent strata of the Bannock Volcanic Member contain biotite, suggesting upper greenschist facies metamorphic conditions. Restoration of 8 km of overlying Phanerozoic miogeoclinal sediment load (Dixon, 1982) at a geothermal gradient of 20 °C/km suggests pressures of 2.6 kb and temperatures of 160 °C, somewhat lower than the 3–6 kb and 400 °C predicted from the greenschist minerals. To the south in northern Utah along the Paris–Putnam–Willard thrust plate, peak Late Cretaceous metamorphic conditions of 4 kb and 300 to 350 °C are suggested (Naeser *et al.* 1983). Several key samples

used in this study come from the Portneuf Narrows area (locality 2). In all other localities, pelitic rocks are uncleaved. They contain mainly white mica ('sericite') and are in lowest greenschist facies.

#### 2.e. Age and correlation

Fossils are extremely rare in the Neoproterozoic rocks of northern Utah and southeastern Idaho. Acritarchs of broadly Neoproterozoic affinity have been found in the Mineral Fork Formation of Utah (Knoll, Blick & Awramik, 1981), but do not tightly constrain the age of the sequence. Coarse-grained quartz sandstones such as those found in the Brigham Group have long been recognized as typical of facies containing the biologically and temporally distinctive Ediacaran fauna (Glaessner, 1984). Unfortunately, despite considerable effort no Ediacaran fossils have been found to date. Early Cambrian trace fossils (*Skolithos*, *Planolites*) are present in the upper part of the Camelback Mountain Quartzite (Fig. 2; Crittenden *et al.* 1971; Oriol & Armstrong, 1971; Link *et al.* 1987), indicating that at least the top of the unit is of Cambrian age.

In the absence of widespread fossils, glacial diamictites have commonly assumed a key role in the correlation of Cordilleran Neoproterozoic strata. Most correlations suggest that the principal diamictite horizons found in northwestern Canada, southern British Columbia and adjacent Washington, Idaho, Utah and California are equivalent and represent the Sturtian glaciation. The Bannock Volcanic Member has traditionally been correlated with mafic volcanic and diamictite-bearing strata from northeast Washington that have been dated using Sm–Nd and Rb–Sr whole-rock analyses at  $762 \pm 44$  Ma ( $2\sigma$ ; Devlin & Bond, 1988), a date consistent with the ages of Neoproterozoic volcanism along the length of the Cordillera. Prior to Aitken's (1991) discovery of a second diamictite in the Rapitan Group of northwestern Canada, the Varanger ice age was thought to be recorded only in high-amplitude sea level changes observed in stratigraphically higher formations, such as the Inkom Formation of Idaho and Utah (Christie-Blick & Levy 1989*b*; Levy, Christie-Blick & Link, in press). The only other radiometric constraint comes from a thin tuff bed in the Brown's Hole Formation in Utah (above the Mutual Formation), which has yielded an  $^{40}\text{Ar}/^{39}\text{Ar}$  age of 580 Ma (recalculated by N. Christie-Blick from Crittenden & Wallace, 1973; no error given; Fig. 2).

Subsidence analysis of other Neoproterozoic to Cambrian successions along the Cordilleran miogeocline has been used to date the initiation of crustal subsidence at approximately  $575 \pm 25$  Ma (assuming a date of 545 Ma for the base of the Cambrian; Armin & Mayer, 1983; Bond, Kominz & Devlin, 1983; Bond *et al.* 1985; Bond & Kominz, 1984; Devlin & Bond,

1988; Christie-Blick & Levy, 1989*a*; Levy & Christie-Blick, 1991*b*). The discrepancy between this date and the age of the initiation of subsidence and rifting implied by the Pocatello Formation and correlative units (perhaps as much as 200 Ma older) was first noted by Link (1984). It has been suggested that there may have been multiple rift events predating continental separation, and that the Pocatello Formation and the lower part of the Brigham Group (at least including strata below the Mutual Formation) were not part of the subsidence cycle that produced the early Palaeozoic Cordilleran miogeocline (Link & Smith, 1992; Link *et al.* 1993). These strata are then interpreted to have been deposited in continental rift basins or intracratonic basin settings (Christie-Blick & Levy, 1989*a*; Levy & Christie-Blick, 1991*a*).

To summarize, the Pocatello Formation and Brigham Group are broadly constrained to be Neoproterozoic in age. Correlations with similar sequences along the western edge of the North American Cordillera suggest that the glacial sediments present near the base of the section were deposited during the Sturtian glaciation, but existing correlations are poorly constrained.

### 3. Methods

Samples were collected from each carbonate unit at Portneuf Narrows, Scout Mountain, Kinport Peak, South Putnam Mountain and Old Tom Mountain south of Pocatello (Fig. 1) in 1991 and 1992 by PKL and LHS. Lehi Hintze (Brigham Young University) generously provided a sample of Blackrock Canyon Limestone from Hardscrabble Creek in the Canyon Range of Utah (Sample HSC; Hardscrabble Creek, Center Sec. 21 T17S R3W; Scipio North 15 minute quadrangle; Millard, 1983). Sampled carbonate units in the Pocatello Formation include the dolomite overlying the upper diamictite and limestones of the uppermost Scout Mountain Member. Other sampled units include the Blackrock Canyon Limestone, and the carbonate member of the Caddy Canyon Quartzite (Figs 1 & 2).

Polished thin and thick sections were prepared for petrographic and cathodoluminescence (CL) examination, respectively. Where possible, non-luminescent (NLM) or moderately-luminescent microspar (MLM) was isolated (5–10 mg) by microdrilling techniques for elemental and isotopic analyses (cf. Kaufman *et al.* 1991). For the determination of Mn and Sr concentrations as well as Mn/Sr in carbonate, aliquots of micro-sampled powders were weighed (typically 1–2 mg) and leached in ultraclean weak (0.5M) acetic acid to avoid the dissolution of clastic components which is noted to occur with the use of even dilute solutions of HCl. After leaching, solutions were decanted and diluted to 100 ml in 2% HNO<sub>3</sub>; residues were dried and weighed to determine per cent

dissolution. Measurements were performed on a VG PQ2+ plasma source mass spectrometer. Gravimetrically determined standards were analysed in order to develop response calibration curves and a 100 ppb <sup>115</sup>In spike was added to the diluted solutions for normalization. Accuracy of elemental abundances determined by this technique are better than ±5% compared to isotope dilution techniques.

Carbon dioxide was evolved from carbonate powders for determination of δ<sup>13</sup>C and δ<sup>18</sup>O by reaction with concentrated phosphoric acid ( $\rho > 1.89$  g/ml) in sealed tubes at 90 °C. In addition, δ<sup>13</sup>C values of total organic carbon (TOC) were determined for each whole-rock sample. TOC was isolated from representative powders (100–200 mg) by repeated acidification and washing; subsequently TOC was combusted (850 °C for 8 h) in sealed Vycor tubes in the presence of cupric oxide. The CO<sub>2</sub> produced from both carbonates and TOC was isolated by cryogenic distillation and measured on a VG PRISM gas source isotope mass spectrometer. Fractionation factors used for the calculation of <sup>18</sup>O abundances of calcite (1.00798) and dolomite (1.00895) were based on analyses of CO<sub>2</sub> prepared at 90 °C. Accuracy of these techniques, as determined by multiple analyses ( $n > 25$ ) of standard materials, is ±0.1‰ for C and ±0.3‰ for O.

Finally, another aliquot of micro-sampled powder was leached in 0.5 M acetic acid for determination of [Rb], [Sr], <sup>87</sup>Rb/<sup>86</sup>Sr and <sup>87</sup>Sr/<sup>86</sup>Sr. After centrifugation the supernate was decanted, dried and 10 ml of 1.5 N HCl was added. A 1% aliquot of the solution was removed and spiked with isotopic tracers for Rb and Sr. Concentrations of these elements were determined by isotope dilution on a Finnigan THQ thermal ionization mass spectrometer. Strontium was isolated from the remaining solution by standard ion exchange techniques and isotopic compositions determined with a Finnigan MAT 262 thermal ionization mass spectrometer with both static and dynamic multi-collection. Interference from <sup>87</sup>Rb was controlled by measuring <sup>85</sup>Rb with an SEM in ion counting mode (cf. Derry, Kaufman & Jacobsen, 1992). Multiple NBS 987 standards analysed during the course of this work yield an average value of  $0.710241 \pm 0.000008$  (uncertainty given as  $2\sigma$  of the mean).

### 4. Results and discussion

Results of all isotopic and elemental analyses are presented in Table 1. It is evident that the isotopic compositions of both C and Sr in the Pocatello Formation and Brigham Group limestones and dolostones exhibit significant stratigraphic variations (Fig. 3). Carbonates of the uppermost Scout Mountain member of the Pocatello Formation are characterized by depleted <sup>13</sup>C abundances (δ<sup>13</sup>C values range from

Table 1. Isotopic analyses of carbonates from the Pocatello Formation and Brigham Group

Sample	Locality	Lithology	Height <sup>a</sup> (m)	$\delta^{13}\text{C}$		$\Delta\delta^c$	TOC <sup>d</sup> (mg C/g)	[Sr] <sup>e</sup> (ppm)	Mn/Sr <sup>e</sup>	$^{87}\text{Rb}/^{86}\text{Sr}^f$	$^{87}\text{Sr}/^{86}\text{Sr}^g$
				$\delta^{13}\text{C}^b$	$\delta^{13}\text{C}_{\text{TOC}}$						
<b>Scout Mountain Member, Pocatello Formation</b>											
67PL90	1	Dolomite w/sparite	0	-3.4	-14.7	-16.7	0.10	69	9.72	—	—
70PL90	2	Dolomite	0	-3.1	-12.9	-22.7	0.07	164	12.62	—	—
91BC06	2	Microspar w/sparite	0	-3.2	-13.7	-20.5	0.11	~150	4.63	—	—
72PL90	2	Microspar w/sparite	90	-5.0	-15.0	-22.7	0.05	264	4.42	—	—
05LHS92	2	Microspar	90	-4.7	-14.8	-22.5	0.08	145	4.37	—	—
73PL90	2	Microspar	110	-4.2	-15.1	-21.9	0.10	1118	0.18	0.00166	0.70811
04LHS92	2	Microspar w/sparite	110	-4.9	-15.4	-17.9	0.06	136	2.72	—	—
03LHS92	2	Microspar	115	-4.7	-15.8	-20.0	0.08	303	2.89	—	—
02LHS92	2	Microspar w/sparite	120	-4.6	-16.2	-22.5	0.27	110	2.94	—	—
01LHS92	2	Microspar w/sparite	125	-4.2	-16.1	n.d.	n.d.	59	1.60	—	—
75PL90	2	Microspar	130	-4.2	-16.1	-24.2	0.24	200	2.17	—	—
<b>Blackrock Canyon Limestone</b>											
07LHS92	3	Microspar	~500	-11.0	-14.1	-21.5	0.16	305	0.36	—	—
08LHS92	3	Sparite	~520	1.1	-13.9	-23.2	0.27	2036	0.01	0.00024	0.70798
09LHS92	3	Microspar w/sparite	~540	-0.2	-15.6	-22.1	0.33	754	0.16	—	—
10LHS92	4	Sparite	~540	0.3	-16.1	-22.4	0.30	533	0.04	0.00011	0.70872
HSC		Sparite		-0.9	-13.4	-25.4	0.32	1057	0.13	0.00064	0.70766
<b>Middle Carbonate Member, Caddy Canyon Quartzite</b>											
69PL90	5	Dolomite	~1800	8.0	-13.4	-21.4	0.94	53	0.78	—	—
91ST06	6	Microspar	~1800	8.8	-17.9	-17.4	0.50	856	0.16	0.00084	0.70885
106PL92	7	Dolomite w/sparite	~1800	3.9	-15.1	-14.5	0.14	406	1.14	—	—

<sup>a</sup> Stratigraphic height above the top of the Upper Diamictite Unit, Scout Mountain Member Pocatello Formation.

<sup>b</sup>  $\delta^{13}\text{C} = [(^{13}\text{C}/^{12}\text{C})_{\text{sample}} / (^{13}\text{C}/^{12}\text{C})_{\text{PDB}}] - 1$ . PDB = Pee Dee Belemnite.  $\delta$  values for other elements are calculated in the same manner.

<sup>c</sup>  $\Delta\delta = \delta^{13}\text{C} - \delta^{13}\text{C}_{\text{TOC}}$ .

<sup>d</sup> TOC = Total organic carbon calculated from yield of  $\text{CO}_2$  after combustion.

<sup>e</sup> Elemental concentrations and ratios determined on a VG PQ2 + PlasmaQuad mass spectrometer.

<sup>f</sup> Determined by isotope dilution techniques on a Finnigan THQ thermal ionization mass spectrometer.

<sup>g</sup> Sr isotope ratios are present day values, not corrected for decay of  $^{87}\text{Rb}$  due to uncertainty in age.

n.d. = not determined.

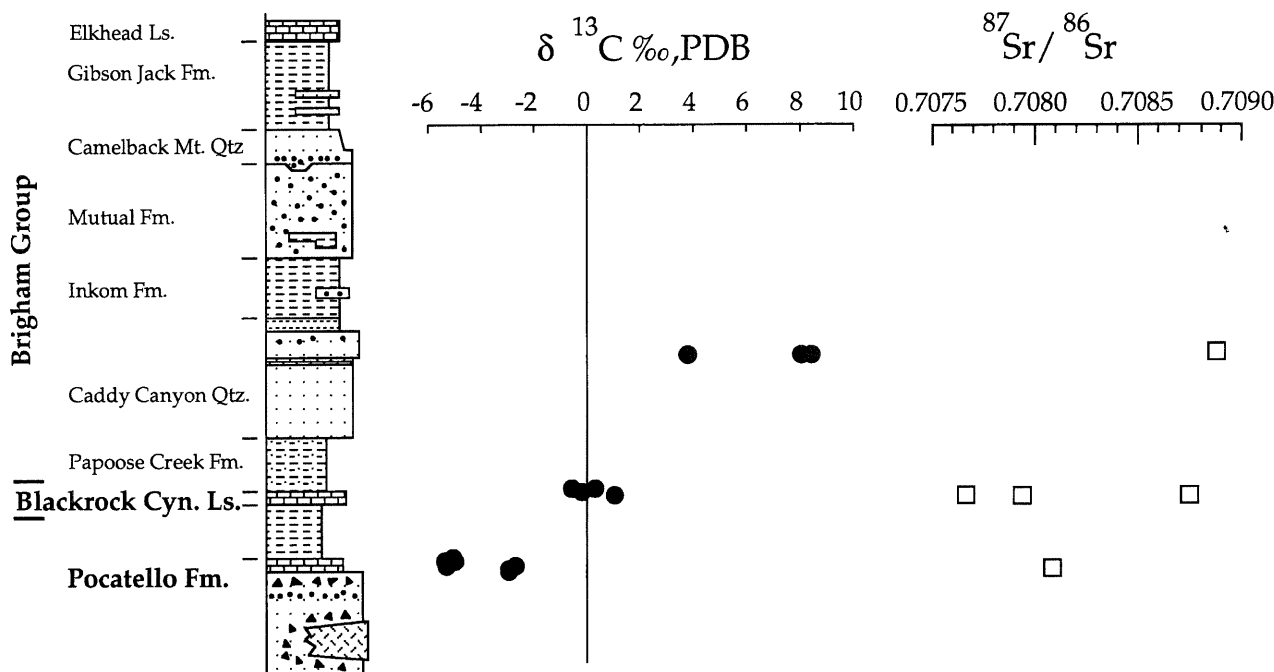


Figure 3. Stratigraphic variation of carbon and strontium isotopes from the Pocatello Formation and Brigham Group. Circles:  $\delta^{13}\text{C}$ . Squares:  $^{87}\text{Sr}/^{86}\text{Sr}$ . Sample 07LHS92 has not been plotted as detailed in the text.

– 5.0 to – 3.1). A single sample from the Blackrock Canyon Limestone yielded a  $\delta^{13}\text{C}$  value of – 11.0; however, all others hover around 0. Three samples from separate localities of the carbonate member of the Caddy Canyon Quartzite give positive  $\delta^{13}\text{C}$  values of 3.9, 8.0 and 8.8. All microsamples contain greater than 95 % carbonate.

Oxygen-isotopic compositions of all samples in this study are significantly depleted in  $^{18}\text{O}$  ( $\delta^{18}\text{O}$  values range between – 13 and – 18) relative to unmetamorphosed and little-altered Neoproterozoic carbonates worldwide, suggesting significant post-depositional alteration. Although  $\delta^{18}\text{O}$  values of these carbonates may have been altered during meteoric diagenesis, the strong depletion in  $^{18}\text{O}$  is likely associated with the exchange of O between carbonate and fluids during greenschist facies metamorphism (Section 2.d).

Total organic carbon (TOC) abundances and isotopic compositions also appear to reflect metamorphic alteration. TOC abundances for all samples are low relative to typical values for unmetamorphosed Neoproterozoic carbonates (Strauss & Moore, 1992; Table 1). In all but two samples, the measured difference in C-isotopic composition between co-existing carbonate and TOC ( $\Delta\delta$ ) is also far below the 28–30‰ commonly recorded for unmetamorphosed Neoproterozoic samples (Table 1; Fig. 4). In a study of Neoproterozoic carbonates from Namibia, Kaufman *et al.* (1991) recorded similarly depleted TOC abundances and  $\Delta\delta$  values for samples from metamorphosed successions. More broadly, a number of authors have documented significant  $^{12}\text{C}$  depletion in partially or entirely graphitized kerogens within

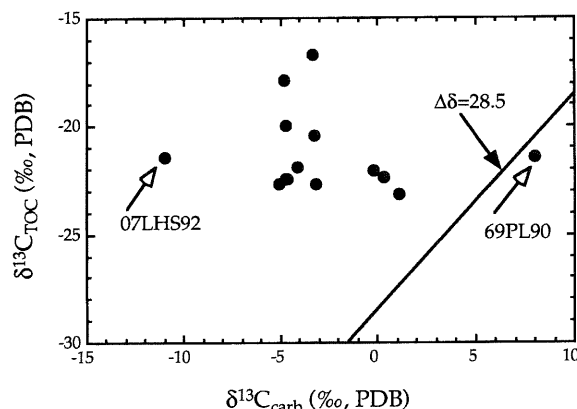


Figure 4. Crossplot of the  $\delta^{13}\text{C}$  values of total organic carbon vs. carbonate carbon. The bold line represents  $\Delta\delta$  (average carbon-isotopic difference between carbonate and TOC) = 28.5, the value observed in a sequence of unmetamorphosed Neoproterozoic sediment from Svalbard and East Greenland (Knoll *et al.* 1986). 07LHS92 – this sample appears to be significantly altered. See Figure 6, Table 1 and discussion in text. 69PL90 – this sample preserves high  $\delta^{13}\text{C}$  values and has a  $\Delta\delta$  value very close to 28.5, suggesting that this is the least-altered sample in the sequence.

metamorphic terranes, with alteration ascribed to the selective loss of  $^{12}\text{C}$ -rich material during thermal maturation and isotopic reequilibration between carbonate and organic carbon (Schidlowski & Aharon, 1992, and references therein).

The very low carbonate  $\delta^{13}\text{C}$  value (– 11.0 – more than 10‰ lower than other samples from the same unit; Fig. 4; Table 1) and low  $\Delta\delta$  value (21.5) from sample 07LHS92 indicate that metamorphic or diagenetic processes have significantly altered this sample. Mass balance considerations make it unlikely that the

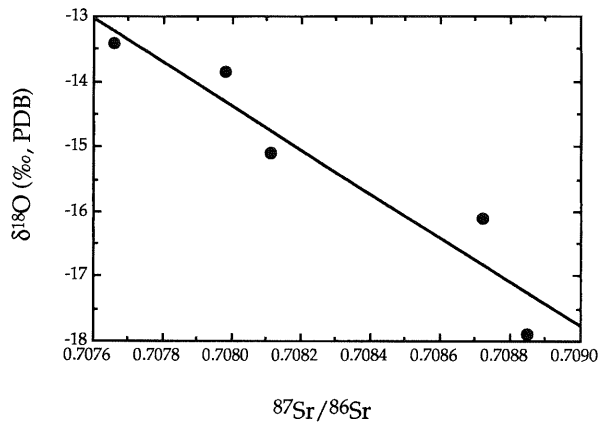


Figure 5. Crossplot of  $\delta^{18}\text{O}$  vs.  $^{87}\text{Sr}/^{86}\text{Sr}$ .  $r^2 = 0.904$ ,  $p > 0.013$ .

diagenetic addition of carbonate via the oxidation of isotopically light organic carbon by itself accounts for the anomaly. Processes such as local decarbonation or the metamorphic reequilibration of C-isotopes between carbonate and organic C are likely alternatives.

Sample 69PL90, which has the highest  $\Delta\delta$  (near the average  $\Delta\delta$  observed for an unmetamorphosed succession of carbonates in Svalbard and East Greenland; Knoll *et al.* 1986) also has the highest TOC and one of the highest  $\delta^{18}\text{O}$  values in the sample set. For these reasons, our confidence in the strong  $\delta^{13}\text{C}$  enrichment noted in this part of the sequence appears to be justified. These analyses add to a growing list of Proterozoic carbonates in which high depositional values of  $\delta^{13}\text{C}$  were preserved despite significant metamorphism (Baker & Fallick, 1989; Wickham & Peters, 1993).

Several samples have both low Mn/Sr ( $< 2$ ) and high Sr concentrations ( $> 500$  ppm) suggesting that  $^{87}\text{Sr}/^{86}\text{Sr}$  representative of Neoproterozoic seawater

may have been preserved (Table 1, Fig. 3; cf. Derry, Kaufman & Jacobsen, 1992).  $^{87}\text{Sr}/^{86}\text{Sr}$  in these samples ranges from 0.7077 to 0.7089. However, in our samples from the Portneuf region, much of the range in  $^{87}\text{Sr}/^{86}\text{Sr}$  is observed within a single unit – the Blackrock Canyon Limestone. Further, there is a strong correlation between  $^{87}\text{Sr}/^{86}\text{Sr}$  and  $\delta^{18}\text{O}$  values (Fig. 5), suggesting that as  $\delta^{18}\text{O}$  was altered to more negative values,  $^{87}\text{Sr}/^{86}\text{Sr}$  increased. If so, the value of 0.70798 determined for sample 08LHS92 is most likely to approximate primary Sr-isotopic compositions for this units in this region. Sample HSC from the Canyon Range of Utah has an even higher  $\delta^{18}\text{O}$  value and lower  $^{87}\text{Sr}/^{86}\text{Sr}$  value (0.70766), which may more closely approximate true  $^{87}\text{Sr}/^{86}\text{Sr}$  values of sea water during deposition of the Blackrock Canyon Limestone.

### 5. Chemostratigraphic correlations

As noted above, the Pocatello Formation diamictites are conventionally interpreted as products of Sturtian glaciation. Strong positive C-isotopic excursions are observed in both post-Sturtian and post-Varanger sequences worldwide (Kaufman & Knoll, 1994), so the overall increase in  $\delta^{13}\text{C}$  from the negative values of the Pocatello Formation to the carbonate member of the Caddy Canyon Quartzite (whose  $\delta^{13}\text{C}$  values are among the most positive observed for Neoproterozoic carbonates) could record the biogeochemical aftermath of either glaciation. On the other hand, previous results lead us to expect that post-Sturtian carbonates should have distinctly lower Sr-isotopic compositions (0.7070 to 0.7075) when compared to those of post-Varanger carbonates (0.7080 to 0.7085; Table 2 and references therein).

Table 2. Comparison of isotopic data for post-Sturtian and post-Varanger carbonate successions

Carbon	Post-Sturtian		Post-Varanger	
	Minimum	Maximum	Minimum	Maximum
Svalbard (b)	-2.5	+8	-4	+5.1 (a)
Namibia (c)	-5	+9.5	-5.5	+6.5
China (d)	—	—	—	+7
Mackenzie Mts (e)	-3.4	+8.3	-4	+7
Australia (f)	-5.6	+9.5	-3	+7
McCoy Creek Group (g)	—	+12	—	—
British Columbia (h)	-5	+9.9	—	—
Idaho (this work)	-5	+8.8	—	—
<b>Strontium</b>				
Svalbard (i)	0.7070–0.7074	—	—	—
Namibia (j)	—	—	0.7080–0.7085	—
Mackenzie Mts (e)	0.7071–0.7072	—	0.7080–0.7085	—
Australia (k)	0.7075	—	—	—
Idaho (this work)	0.7076 or lower	—	—	—
E. Humboldt Range (l)	0.7073–0.7076	—	—	—

(a) This section is truncated; (b) Knoll *et al.* 1986; Fairchild & Spiro, 1987; (c) Kaufman *et al.* 1991; (d) Lambert *et al.* 1987; (e) Narbonne, Kaufman & Knoll, 1994; (f) Jenkins *et al.* 1992; Jenkins, 1992; Walter *et al.* 1992; (g) Wickham & Peters, 1993; (h) Ghent & O'Neil, 1985; (i) Derry, Kaufman & Jacobsen, 1992; (j) Kaufman, Jacobsen & Knoll, 1994; (k) Veizer *et al.* 1983; (l) M. Peters & S. Wickham, pers. comm. 1993.



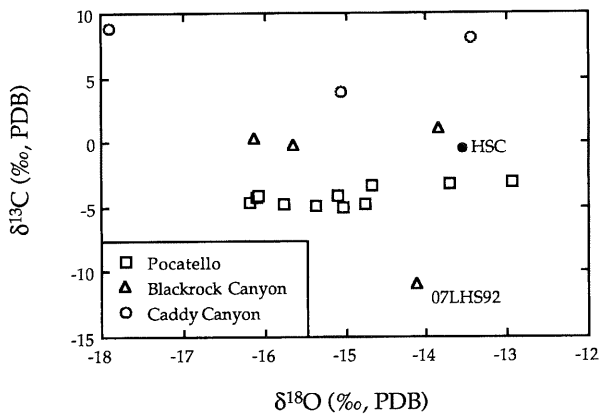


Figure 6. Crossplot of  $\delta^{13}\text{C}$  versus  $\delta^{18}\text{O}$ . There is no significant positive correlation between  $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$  within any of the units. 07LHS92 – this sample falls well outside the trend observed for other samples from the Blackrock Canyon Limestone, suggesting post-depositional alteration. HSC – Blackrock Canyon Limestone from Hardscrabble Canyon (provided by L. Hintze). Christie-Blick (1982) assigns this unit to the carbonate member of the Caddy Canyon Quartzite.

### 5.a. The carbon isotope record

Crossplots of  $\delta^{13}\text{C}$  versus  $\delta^{18}\text{O}$  are commonly employed in studies of diagenesis (e.g. Fairchild, Marshall & Bertrand-Sarfati, 1990), as meteoric alteration commonly results in the depletion of  $^{18}\text{O}$  or both  $^{18}\text{O}$  and  $^{13}\text{C}$  in carbonates. As shown in Figure 6,  $\delta^{13}\text{C}$  values vary little with decreasing  $\delta^{18}\text{O}$  values within individual horizons of the Pocatello Formation and Blackrock Canyon Limestone. Retention of near-depositional C-isotopic composition is common, albeit not universal, in both Neoproterozoic limestones and dolostones (reviewed in Kaufman & Knoll, 1994).

Stratigraphic excursions comparable to those in this succession have been recorded elsewhere in the Cordillera (Table 2). In the Windermere Supergroup of the Mackenzie Mountains, northwestern Canada,  $\delta^{13}\text{C}$  values range from  $-4.6$  in cap limestone and dolostone just above the Varanger-aged Ice Brook diamictites to  $+7.0$  in the upper part of the overlying Sheepbed Formation (Narbonne, Kaufman & Knoll, 1994). Limited isotopic data on units below the Ice Brook Formation document a similar excursion from the Sturtian-aged Rapitan Group, which includes both diamictite and iron-formation, to the Keele Formation, which underlies the Ice Brook diamictites. Klein & Beukes (1993) reported  $\delta^{13}\text{C}$  values as low as  $-3.4$  in carbonates within the upper Rapitan Group, whereas Narbonne, Kaufman & Knoll (1994) recorded  $\delta^{13}\text{C}$  values as high as  $+10.1$  in Keele carbonates. Keele limestones yield  $^{87}\text{Sr}/^{86}\text{Sr}$  values between 0.7071 and 0.7072 (Kaufman, Jacobsen & Knoll, 1994).

Elsewhere in the Cordillera similar excursions have been documented in Neoproterozoic successions, but their interpretation as post-Sturtian or post-Varanger

remains conjectural. Ghent & O'Neil (1985) reported  $\delta^{13}\text{C}$  values ranging between  $-5.0$  and  $+9.9$  from Neoproterozoic marbles in southeastern British Columbia. Although the stratigraphic context has been somewhat obscured by folding, the strata are underlain by the Irene Volcanics and Toby Conglomerate (Poulton & Simony, 1980). The Toby Conglomerate is generally thought to be glacial in origin, and its sedimentary and tectonic setting is similar to that of the Pocatello Formation (Aalto, 1971). Wickham & Peters (1993) reported  $\delta^{13}\text{C}$  values of  $+6.5$  to  $+9.9$ , with an extreme as high as  $+12$ , in marbles of the Neoproterozoic McCoy Creek Group exposed in the East Humboldt Range, Schell Creek Range and Pilot Range (upper amphibolite facies terranes). As remarked upon by these authors, such anomalously positive  $\delta^{13}\text{C}$  values are so striking that they form a distinctive regional stratigraphic marker. In concert with lithostratigraphic similarities, the extreme  $^{13}\text{C}$  enrichments suggest that the British Columbia, McCoy Creek, and Caddy Canyon carbonates are correlative as first noted by Crittenden *et al.* (1971; see also Peters, Wickham & Miller, 1992).

Given the hypothesis of a late Neoproterozoic supercontinent juxtaposing western North America and Australia, comparison of our data with values from Australia is warranted. Jenkins *et al.* (1992) report preliminary C-isotopic data from Vendian successions in the Adelaide Geosyncline and Amadeus Basin that compare closely with the curve for the post-Varanger succession in the Mackenzie Mountains (Narbonne, Kaufman & Knoll, 1994). In the Flinders Ranges,  $\delta^{13}\text{C}$  increases stratigraphically from  $-3$  in the Nuccaleena Formation (Williams, 1979) to as high as  $+7$ . A cap dolomite on top of the Sturtian Landrigan Tillite of the Kimberley region yielded a  $\delta^{13}\text{C}$  value of  $-5.6$  (Williams, 1979); higher in the post-Sturtian/pre-Marinoan succession of the Adelaide Geosyncline, Veizer & Hoefs (1976) reported a  $\delta^{13}\text{C}$  value of  $+9.5$  from the Elatina Formation.

Other regions also show C-isotopic excursions in association with both Sturtian and Varanger glaciations. For example, in Svalbard and East Greenland, carbonates overlying Varanger tillites are depleted in  $^{13}\text{C}$  ( $\delta^{13}\text{C} \sim -2$  to  $-4$ ). Higher in the incomplete post-Varanger succession  $\delta^{13}\text{C}$  values rise to  $+5.1$  (Knoll *et al.* 1986; Fairchild & Spiro, 1987). A positive isotopic excursion from  $-2.5$  to  $+8.0$  lower in the succession is thought to record the biogeochemical aftermath of the Sturtian glaciation (Knoll *et al.* 1986). Probable post-Sturtian and post-Varanger excursions in Namibia show trends from  $-5.0$  to  $+9.5$  and from  $-5.5$  to  $+6.5$ , respectively (Kaufman *et al.* 1991). In the Neoproterozoic Bambui Group of Brazil, carbonates that overlie diamictites yield  $\delta^{13}\text{C}$  values as high as  $+10$  to  $+16$ , but the stratigraphic level of the anomaly is poorly constrained (Iyer, Krouse & Babinski, 1992; Iyer *et al.* in press).

In summary, data from several continents indicate that strong carbon-isotopic excursions like that recorded in the Pocatello/Brigham succession occurred during at least two intervals of the late Neoproterozoic Era. However, in every case where post-Sturtian and post-Varanger excursions are known from a single region, post-Sturtian positive excursions peak at  $\delta^{13}\text{C}$  values of +8.0 to +9.5 (or higher), while no known post-Varanger peak exceeds +7 (Table 2). This difference suggests that the Pocatello/lower Brigham succession is post-Sturtian and pre-Varanger in age, corroborating earlier lithostratigraphic correlations.

### 5.b. The strontium isotope record

There is a considerable spread in the values of  $^{87}\text{Sr}/^{86}\text{Sr}$  both within and among stratigraphic units of the Pocatello Formation and Brigham Group. Although there is nothing in the Sr concentration, Mn/Sr, or  $^{87}\text{Rb}/^{86}\text{Sr}$  of the measured samples that raises suspicion or allows us to identify individual samples as altered, the variation within a single horizon indicates that variable alteration has occurred. As noted above, the correlation between  $\delta^{18}\text{O}$  values and  $^{87}\text{Sr}/^{86}\text{Sr}$  values also suggests that Sr-isotopic compositions have been altered by metamorphic fluids. Oxygen is likely to be the most sensitive indicator of fluids that are able to influence minor components such as Sr. Alteration by meteoric and metamorphic fluids tends to increase  $^{87}\text{Sr}/^{86}\text{Sr}$  in carbonates, which is consistent with the correlation between  $\delta^{18}\text{O}$  and  $^{87}\text{Sr}/^{86}\text{Sr}$ . Therefore, among samples collected at our principal localities in the Portneuf Narrows region, sample 08LHS92 may most nearly approach the Sr-isotopic composition of seawater at the time of deposition. A correlative sample from a distant locality – sample HSC from the Canyon Range of Utah – has lower  $^{87}\text{Sr}/^{86}\text{Sr}$  and higher  $\delta^{18}\text{O}$  values than other samples analysed in this study, suggesting that this sample comes still closer to approximating seawater  $^{87}\text{Sr}/^{86}\text{Sr}$  during the time of deposition of the Blackrock Canyon Limestone. If the McCoy Creek and Pocatello/Brigham successions are correlative, then their Sr-isotopic compositions should be comparable. Recent analyses of the Sr-isotopic composition of McCoy Creek marbles in the East Humboldt Range indicate  $^{87}\text{Sr}/^{86}\text{Sr}$  values between 0.7073 and 0.7076 (Table 2; M. Peters & S. Wickham, pers. comm., 1993).

Although only a limited number of post-Sturtian and pre-Varanger  $^{87}\text{Sr}/^{86}\text{Sr}$  values have been reported, values between 0.7073–0.7077 are consistent with those observed in other post-Sturtian successions. In Spitsbergen, Derry *et al.* (1989) reported post-Sturtian values as high as 0.7074 in carbonates that record post-Sturtian  $^{13}\text{C}$  enrichment. Veizer *et al.* (1983) reported a single  $^{87}\text{Sr}/^{86}\text{Sr}$  value of 0.7075 for the

Brighton Limestone of the Adelaide Geosyncline, South Australia. The Brighton Limestone is a partial correlative of the Elatina Formation, from which the post-Sturtian  $\delta^{13}\text{C}$  value of +9.5 was obtained. In contrast, post-Varanger carbonates in which  $\delta^{13}\text{C}$  is markedly positive yield  $^{87}\text{Sr}/^{86}\text{Sr}$  values of 0.7080 to 0.7085 (Table 2; Kaufman, Jacobsen & Knoll, 1994). Thus, the Sr-isotopic data corroborate C-isotopic results in suggesting a post-Sturtian and pre-Varanger age for the upper Pocatello and lower Brigham succession.

### 6. Conclusions

The Pocatello and Brigham succession illustrates both the potential and problems of applying isotope chemostratigraphy to predominantly siliciclastic Neoproterozoic successions. Because samples are necessarily few, it can be difficult to determine whether or not geochemical signatures are strongly influenced by diagenesis. Interpretation is easier for carbon, where inorganic and organic phases offer two views of secular trends. For strontium, however, evaluation of a small number of samples can be problematic. While geochemical screening can eliminate obviously altered samples, confidence in interpretation comes from a large and consistent data set. During part of the Neoproterozoic Era, the  $^{87}\text{Sr}/^{86}\text{Sr}$  of sea water was so low that unaltered values can be recognized on the basis of limited analyses, but when Sr-isotopic ratios fall in the range of 0.7075 to 0.7085, differentiation between an unaltered 570 Ma old limestone and a moderately altered 700 Ma old sample can be impossible in the absence of additional data.

A further problem concerns the density of existing global data. For the Vendian interval, detailed studies of C and Sr chemostratigraphy that are well integrated with lithostratigraphic, biostratigraphic, petrological, and other geochemical data have been completed for successions in North America, Europe, Siberia, China, Africa, the Middle East, and Australia. The results are mutually consistent and provide a reliable framework for chemostratigraphic correlation (Knoll & Walter, 1992; Kaufman & Knoll, 1994; Kaufman, Jacobsen & Knoll, 1994). For the earlier Neoproterozoic, basal studies are fewer and our understanding of secular variation is correspondingly more impressionistic, although there is little reason to doubt the general nature and magnitude of recorded events. Particularly for  $^{87}\text{Sr}/^{86}\text{Sr}$ , the limited number of samples analysed to date prevents us from being absolutely certain that seawater  $^{87}\text{Sr}/^{86}\text{Sr}$  never reached values above 0.7080 between the Sturtian and Varanger ice ages.

Despite these problems, we believe that isotopic chemostratigraphy provides a useful new perspective on Neoproterozoic correlation in western North America. In particular, recognition that the magnitude

of the positive  $\delta^{13}\text{C}$  excursion following the Sturtian glaciation is uniquely large provides us with the potential for identifying this interval even when carbonates are few and thin. Our data support earlier lithostratigraphic correlation of the Pocatello and related diamictites in the northwestern United States and British Columbia with the older (Rapitan) diamictite succession in the Mackenzie Mountains. This is consistent with the sequence stratigraphic correlations of Christie-Blick & Levy (1989a) and Levy, Christie-Blick & Link (in press) in which a series of incised valleys within the Inkom Formation are interpreted to represent sea-level fluctuations produced by the Ice Brook glaciation to the north.

The correlation of predominantly carbonate with predominantly siliciclastic sequences has always been one of the most difficult problems of stratigraphy, regardless of the age of the rocks. Multiple lines of evidence are necessary to discriminate among alternative correlations with radically different geological consequences. Although the Neoproterozoic Pocatello Formation and Brigham Group are mainly siliciclastic, the presence of relatively thin carbonate units has enabled us to provide additional support for the conventional correlation of these strata and other post-Sturtian successions.

Chemostratigraphic correlation of siliciclastic Neoproterozoic sequences will always require a high degree of caution. The relatively small number of available samples makes it difficult to apply rigid criteria for the screening of isotopic data, and may hamper the identification of diagenetic effects. Fortunately, our knowledge of the effects of metamorphism and diagenesis on the isotopic and chemical composition of carbonate-dominated sequences (Wickham & Peters, 1993, and references therein) allows us to suggest reasonable interpretations of limited isotopic data sets. In addition, new chemostratigraphic tools promise to increase our ability to correlate Neoproterozoic siliciclastic rocks. Specifically, Ross, Bloch & Krouse (1991) have shown that the isotopic composition of sulfide sulfur underwent a dramatic shift during the late Neoproterozoic. Although the technique requires data sets considerably larger than considered here, this isotopic system may hold unique promise for the chemostratigraphic analysis of Neoproterozoic siliciclastic sequences.

In unaltered successions, stratigraphic variations in  $\delta^{13}\text{C}$  of TOC parallel those observed in carbonate carbon. Therefore, isotopic analyses of organic carbon preserved in siliciclastic sediments provide a possible proxy for the carbonate signal. Unfortunately, like that of carbonate carbon, the isotopic composition of organic carbon can be altered syndepositionally by bacterial reworking or later, during burial diagenesis and metamorphism. The H/C ratio can potentially identify the latter class of alteration, but offers little hope of identifying the former. Thus, in the absence of

data on co-existing carbonates, there are at present no robust checks on the isotopic alteration of sedimentary organic matter. Extreme caution must, therefore, be exercised in employing the C-isotopic composition of TOC or kerogen as a correlation tool. Parallels in the shape of the curve of variation in  $\delta^{13}\text{C}$  of TOC may mirror those seen in the record of carbonate isotopic variation, but as can be seen from Table 2, differences in as little as 2‰ may be key in discriminating among distinct stratigraphic events. Variation of this magnitude is easily introduced into the record of TOC by post-depositional alteration.

In conclusion, we suggest that integrative approaches hold the best promise for stratigraphic correlation of unfossiliferous siliciclastic Neoproterozoic successions. Viewed with appropriate caution, the isotopic composition of thin carbonate sequences can be used as a key to unlock difficult stratigraphic problems. At the very least, chemostratigraphy allows the formulation of hypotheses which may be scrutinized using other techniques.

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