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Platform drowning leading to cool-water carbonate deposition: evolution of a Late Ordovician (Turinian-Chatfieldian) mixed-sediment platform within the Taconic orogen (Long Point Group, Newfoundland Appalachians)

George R. Dix and Elliott T. Burden

G.R. Dix. Ottawa–Carleton Geoscience Centre, and Department of Earth Sciences, Carleton University, 2203 Herzberg Laboratories, Ottawa, ON, Canada, K1S 5B6;

E.T. Burden. Department of Earth Sciences, Memorial University of Newfoundland, St. John’s, NL, Canada, A1B 3X5

Corresponding Author: George R. Dix (email: george.dix@carleton.ca)
ABSTRACT
Late Ordovician (Turinian-Chatfieldian) drowning of a mixed carbonate-siliciclastic platform within the Taconic Orogen (Newfoundland Appalachians) is recorded by net deepening of an initial warm, shallow-water platform succession (Lourdes Formation) culminating in a metre-scale-thick condensed interval that characterizes a drowning succession punctuated by storm deposits. Composition of transported material suggests that seaward drowning was coupled with back-stepping of a high-energy carbonate factory related to hinterland uplift and erosion that would eventually lead to drowning of the outer platform beneath marine-transported siliciclastic sediments (Winterhouse Formation). In the new offshore shelf setting, a sparse reciprocal stratigraphy of fine- to very coarse-grained phosphatic carbonate and mixed sediment is interpreted to document gravity-flow deposition downgradient from either a sustained or episodically developed high-energy cool-water carbonate source along the inner shelf. Transported carbonate was cemented rapidly at temperatures no warmer than 16 to 23°C, possibly within a seasonal oceanic thermocline. An upsection decrease in abundance of carbonate by the early Edenian is associated with dramatic increase in siliciclastic supply. The Turinian-Edenian succession of platform drowning, oceanographic transition to cool-water carbonate production, and, later, its termination by increased siliciclastic supply reflects a first-order tectonic control proximal to uplift within the Taconic Orogen. Similar structural and oceanographic changes along the contemporary distal Laurentian margin provides the basis, with improved biostratigraphic control, for future analysis of the significance of proximal-distal stratigraphies in response to regional foreland tectonism.

Late Ordovician, platform drowning, cool-water carbonates, tectonism, reciprocal stratigraphy
Introduction

Drowning of a shallow-water carbonate platform occurs when the rates of carbonate production and-or accumulation fall below the rate of relative rise in sea level (Schlager, 1989; Kim et al., 2012). Change in water depth, however, can trigger numerous other factors of a more immediate environmental impact on carbonate productivity: for example, nutrient loading, seafloor hypoxia, salinity, change in depositional loci related to modification of current paths and wave energy, and marine burial by prograding siliciclastics (Schlager, 1989; see Table 1 and references in Erlich et al., 1990; John and Mutti, 2005; Betzler et al., 2013). The physical expression of drowning may be an abrupt onlap that defines the drowning unconformity (Schlager and Camber, 1986), a condensed interval (e.g., Bernoulli and Jenkyns, 1974; Szulczewski et al., 1996), burial by marine-transported siliciclastics (Schlager, 1989; Erlich et al., 1990), or combinations of the above (Gold et al., 2017).

The Upper Ordovician (Turinian-Edenian) Long Point Group (Bergström et al., 1974) in the Newfoundland Appalachians (Fig. 1) characterizes a transgressive (carbonate platform)-regressive (siliciclastic shelf) depositional cycle within the Taconic Orogen (Quinn et al., 1999). The deepening-upward platform succession is succeeded by a reciprocal limestone-siliciclastic stratigraphy in the lowermost shelf succession. This latter stratigraphic fabric was interpreted previously as a product of storm reworking of the older platform (Quinn et al., 1999). The purpose of our study is to illustrate how this stratigraphic fabric is part of a longer term history of platform aggradation, step-back, and eventual demise in response to tectonic and sediment supply controls strongly influential in tectonically active basins (e.g., Mullins et al., 1991; Galewsky et al., 1996; Wilson, 2000; Mernier et al., 2014). The reciprocal stratigraphy reflects oceanographic transition from warm- to cool-water carbonates as a byproduct of subsidence.
Geologic Framework

The Long Point Group forms a narrow outcrop belt along the western shore of the Port au Port Peninsula, western Newfoundland (Fig. 1). Its basin developed within the Taconic Orogen following collisional tectonics that caused foundering of a Middle Ordovician foreland carbonate platform (Table Head Group) that was buried beneath orogen-derived siliciclastics of the Goose Tickle Group and associated transported thrust slices (e.g. Humber Arm allochthon, Fig. 1) (Stenzel et al., 1990; Waldron et al., 1993; Quinn, 1995; Zagorevski and van Staal, 2011).

Allochthon transport was reactivated during the Acadian Orogeny, creating west-dipping to locally overturned Long Point strata (Fig. 1; Waldron et al., 1993; Stockmal et al., 2004).

Three constituent formations of the Long Point Group (Fig. 1) characterize a transgressive-regressive depositional cycle (Quinn et al., 1999). A relatively thin (~80 m) mixed-sediment platform succession, the Lourdes Formation, consists of a disconformable superposition of three members that record net deepening (Fig. 2A; Stait and Barnes, 1991), with orogen-derived siliciclastic sediment capping bounding sub-aerial disconformities (Batten Hender and Dix, 2008). The Winterhouse Formation is ~900 metres thick with exposure of its lower ~200 metres with sparsely distributed limestone beds in a siliciclastic succession (Loc. 1 and 2; Fig. 1), and exposure of the upper ~10 metres of sandstone and shale in conformable contact with the Misty Point Formation (Loc. 4b, Fig. 1; Quinn et al., 1999). The intervening covered section is likely siliciclastic as defined by an equivalent section in a petroleum well, ~20 km to the southwest (Newfoundland Hunt Oil Co, 1996). The Winterhouse-Misty Point section characterizes shallowing from shelf to shoreface and fluvial facies, respectively (Gillespie, 1998; Quinn et al., 1999). The Misty Point Formation is faulted against lower Devonian strata (Burden et al., 2002).
For biostratigraphic age control of the Lourdes Formation, no faunal group establishes a definitive zonal assignment. Stait and Barnes (1991) noted, however, that the base of the formation is late Chazyan or early Turinian in age; and, the interval from the Shore Point Member (unit SP3; Fig. 2A) to the top of the formation could be restricted (based on correlation of nautiloid species with Laurentian stratigraphy) to the Blackriveran (or Turinian), or a broader Blackriveran to Rocklandian (early Chatfieldian) range incorporating other shelly biota.

There is no definite zonal assignment for the lower 54 metres of the Winterhouse Formation (Quinn et al., 1999). From 54 to 185 metres (Loc 1 and 2; Fig. 1) a late Chatfieldian (Shermanian) age, within the *C. spiniferus* Zone, might be interpreted (Fig. 2B). However, Quinn et al. (1999) described a graptolite species ~ 84 to 144 metres above the formation base that resembles *Amplexograptus maxwellii*, a species ranging through the Mohawkian (Fig. 2B; Bergstrom and Mitchell, 1986; Goldman et al., 2002). Thus, an older Chatfieldian age cannot be excluded for this interval. Graptolites defining the lower *G. pygmaeus* Zone, and an early Cincinnatian (Edenian) age, occur in the uppermost (> 185 metres) lower Winterhouse Formation (Loc. 2), and in the upper 10 metres of the formation (Loc. 4b; Quinn et al., 1999).

Three acritarch assemblage zones are represented in the lower Winterhouse succession, and define a mid- to late Caradocian age assignment but with uncertainty of how much of the Chatfieldian is represented (Gillespie, 1998). Of potential age significance, Gillespie (1998) reported an elevated abundance of *Gloeocapsamorpha* sp. in the lower 5 metres of the Winterhouse Formation (Gillespie, 1998). In Laurentian platform successions, elevated abundances of *G. prisca* are associated with the Guttenberg Isotope Carbon Anomaly (GICE) within the lower Chatfieldian (Jacobson et al., 1995; Ludvigson et al., 1996).
Analytical Methods

A composite section of the uppermost Lourdes (upper Beach Point Member) and lower Winterhouse formations was mapped incorporating GPS coordinates, extending south from Long Point (Locs. 1 and 2; Fig. 1). Thin sections of the Lourdes Formation (Batten Hender, 2007) were examined, and additional thin sections were stained for calcite, dolomite, and potassium feldspar (Rosenblum, 1956; Dickson, 1966). In addition to standard (transmitted, reflected) microscopy, cathodoluminescence of polished thin sections was examined using an ELM-2 luminoscope operating at 0.5 milliamps, 12 kV, and < 40 millitorr; and, epifluorescence was examined using a Leica Laborlux fluorescence microscope with a 100 W mercury lamp and an excitation filter of 515-560 nm. Ultrastructure and qualitative geochemistry were assessed using a Tescan Vega-II XMU VP-SEM (Carleton University). Trace element analyses were based on rastered (10x10 micron) areas analyzed by an electron microprobe (Carleton University); operating conditions of the automated 4-spectrometer Camebax MBX electron probe were 15 kV and a 20 nano-ampere beam current, using a wavelength dispersive x-ray analyzer. Calcite powder was obtained by drilling thin-section cutoffs, and split for stable carbon and oxygen isotope analysis (Queen’s University) and Sr$^{87}$/Sr$^{86}$ (Carleton University) analyses. For the former, CO$_2$ involved bathing samples in 100% H$_3$PO$_4$ with 2 hours of reaction at 258°C; samples were run on the facilities DELTA$^{+}$plusXP mass spectrometer. Sr$^{87}$/Sr$^{86}$ ratios of calcite were determined by standard procedure, see [http://iggrc.carleton.ca/info-clients/analytical-information](http://iggrc.carleton.ca/info-clients/analytical-information).
Facies Types and Associations

Beach Point Member (upper Lourdes Formation)

Facies types and associations

The upper few metres of the Beach Point Member (Fig. 3A) are defined as a new informal unit, BP3 (Fig. 2A), relative to the previous subdivision of the Beach Point Member, BP1 and BP2 (Stait and Barnes, 1991). The Beach Point Member consists mostly of coated-grain-bearing skeletal grainstone and packstone, nodular silty skeletal packstone, siltstone, and minor calcareous sandstone (Table 1 and Fig. 2A; Stait and Barnes, 1991; Batten Hender and Dix, 2008). Unit BP3 disconformably overlies this succession and, itself, is overlain abruptly by siliciclastics of the Winterhouse Formation (Fig. 3B). The upper boundary is ~30 cm above the last appearance of an in situ colony of Labyrinthites chidlensis, the characteristic coral of the Lourdes Formation (James and Cuffey, 1989).

Unit BP3 contains a very thin bedded and nodular fabric (Fig. 3A,C) consisting of calcareous siltstone, siliciclastic-bearing lime mudstone, and sandy skeletal-lithoclastic grainstone (facies BP3a-c, Table 1). Grainstone forms discontinuous beds of variable thickness (Fig. 3C) due to top–down bioturbation. Erosional surfaces are numerous (facies BP3d) and of two types: (1) well burrowed, gently undulating paleosurfaces with mm- to cm-scale paleotopography developed on lime mudstone (Fig. 3D); and (2) more irregular, jagged paleosurfaces developed on grainstone, and locally blackened due to pyritization. Scattered across both surfaces are heavily abraded, fragmented large-body fossils (Table 1), and individual to clustered gravel- to cobble-size lithoclasts of grainstone and lime mudstone (Fig. 3E) similar to the underlying facies. Many clasts contain borings similar to Trypanites (Fig. 3E). Some are differentially blackened due to pyrite, others display substantial pyrite encrustation (Fig. 3F).
In thin section, silt to sand-size allochems of facies BP3a-c, but in particular grainstone, include sand- to gravel-size clasts of blackened (pyritic) silty lime mudstone (Fig. 4A), fragmented and strongly abraded skeletal material (Table 1), rare phosphatic grains (shells, and rounded sand-size grains of indeterminate origin), and fragments of delicate (floc-like) to more robust rounded fragments of calcified cyanobacteria. The latter are represented by *Girvanella*, the helically coiled *Obruchevella* (Fig. 4B; Burzin, 1995), clustered and encrusting coccoid forms (Fig. 4C), and micritic calcispheres. The grainstone facies also includes silt- to sand-size rounded to very angular silicate grains (quartz, potassium feldspar, rare chloritized serpentine), rounded to angular detrital Fe-zoned dolomite (Fig. 4D), and rare angular gravel-size lithoclasts of feldspathic arenite cemented by Fe-zoned dolomite of similar size as detrital forms.

Lime mudstone (facies BP3b) is a calcisiltite (Flügel, 2010) consisting of a mosaic of rounded to angular detrital calcite of fine-grained silt (< 20 microns) size, and angular siliciclastic grains of similar size (Fig. 4E). There are also two very coarse-grained (gravel, cobble) detrital facies in unit BP3 (Fig. 2A). A matrix- to framework-supported conglomerate (facies BP3e: Fig. 4F; Table 1) contains well rounded but tabular, differentially blackened (pyritic), lithoclasts of lithologies similar to facies BP3b, c, in addition to a rounded pebble-size clast lithic arenite with fragments of chloritized serpentine. The matrix is a mixed sandy grainstone similar to facies BP3c, and contains detrital dolomite as above. A second coarse-grained facies (BP3f; Table 1) is represented by a matrix-supported lithoclastic breccia that forms the uppermost 10-15 cm of the formation (Figs. 3B). Clasts are very irregular, angular, pebble to cobble in size, and randomly to locally vertically arranged. All clasts display blackened (pyritic) margins, including inside margins of *Trypanites* borings, except where a clast was fractured during or prior to deposition (Fig. 4G). Clast lithologies suggest reworking of underlying mudstone and grainstone facies in a
matrix that includes rare fragments of *Nuia*, calcitized fragments of the calcareous alga *Vermiporella*, sand-size grains of chloritized serpentine, and rare phosphatic shell material.

Environmental Interpretation

Unit BP3 has depositional and syndepositional diagenetic attributes in common with condensed intervals (Gómez and Fernández–López, 1994). The thin stratigraphic fabric defines high-order variation in deposition, mechanical (erosion, skeletal breakage) and biological (bioturbation) reworking of the seafloor. Erosional surfaces and reworked bored clasts illustrate formation and modification of hardgrounds, illustrating similar attributes found in other Upper Ordovician successions (Brett and Brookfield, 1984) and in general (MacEachern et al. 2007).

Within this succession, beds of calcisiltite, skeletal grainstone, and conglomerate are interpreted to define detrital facies along an increasing energy gradient of current or wave activity likely defining periods of storm transport and seafloor reworking across this sediment-starved platform. Siltstone forms a background lithology, as it does in the underlying parts of the Beach Point Member (Stait and Barnes, 1991; Batten Hender and Dix, 2008). Thus, detrital crinoid debris, in combination with reworked calcified bacteria and dasycladalean algae, identify a carbonate source beyond the plane of the outcrop belt that hosted a normal marine phototrophic benthic community (Mamet et al., 1984; Riding, 1991; Burzin, 1995, 1996). Calcisiltite, with comminuted skeletal and-or lithoclastic material, likely accumulated from plumes of carbonate mud stirred up during storms, as recognized along modern platforms (*e.g.*, Acker et al., 2002). Most lithoclasts appear to be locally derived, but lithic arenite and detrital dolomite identify exotic sources to the upper Beach Point Member, and illustrating erosion of older strata (see below). Water depth for both the carbonate source and unit BP3 remains uncertain. Ordovician
dasycladalean algae occupied deeper open-shelf settings than found today (Mamet et al., 1984); and, along Vendian and Cambrian platforms, *Obruchevella* is associated with environments of reduced accumulation rates, or deeper-water (mid- to outer-ramp) neritic environments (Burzin, 1996; Elicki, 1999).

Differentially blackened (pyritized) skeletal grains and lithoclasts indicate formation of sulphide under anoxic conditions. With evidence for reworking of the ancient seafloor, a process combining shallow burial, then exhumation (e.g., Baird and Brett, 1986; Taylor and Wilson, 2003) may suitably explain reworked pyritized clasts. The presence of non-pyritic fracture surfaces (Fig. 5A) suggests that pyritization predates final burial. This process coupled with strong reworking into skeletal sand, then cementation, defines the final state of the Lourdes platform prior to influx of Winterhouse siliciclastic sediment (Fig. 3B).

**Lower Winterhouse (Siliciclastic Succession)**

*Facies and Facies Associations*

Quinn et al. (1999) provided a general description and stratigraphic log showing 320 metres of exposed lower Winterhouse Formation extending from Loc. 1 through 2 (Fig. 1). It is likely that this thickness includes repetition across faults. The majority of the section consists of calcareous siltstone (our facies *W1*) finely interbedded with planar bedded and laminated fine-grained calcareous feldspathic arenite (our facies *W2*, Table 2; Fig. 5, 6A, Table 2). These authors noted that sandstone beds display erosional bases; some bedding tops display asymmetric ripples, and Reynolds (2015) illustrates interference ripples; local amalgamated bedding is developed, thinning along strike; hummocky cross-stratification (including a decimeter-scale paleosurface morphology) appears more prevalent in the lower section; and, paleocurrent indicators, though
not abundant, indicate southerly transport. Skeletal material in the siliciclastic succession is of marine origin (Table 2), generally fragmented and dispersed but Quinn et al. (1999) reported basal lags in some beds. They also noted bioturbation concentrated along the tops of sandstone beds, and subsequent analysis has revealed a relatively diverse ichnofossil suite, including *Paleodictyon* sp. and sub-vertical tubular burrow systems filled with siltstone (Reynolds, 2015; Reynolds and McIlroy, 2017).

We add new details to the above. Some sandstone beds contain silt to medium sand-size detrital dolomite similar to that in unit BP3 of the Lourdes Formation. Bed texture and thickness are not random, but define an upsection variation in fining/thinning– and coarsening/thickening-upward bedset succession (Fig. 5). Bed-top waveforms include interference ripples (Fig. 6B) with *Chondrites* burrows, and hummocky cross-stratification locally developed above 170 m above the formation base (Fig. 5). Bidirectional current lineations, though not common, occur throughout the section and identify a predominant NE–SW orientation and a subordinant NW–SE orientation (Fig. 5 inset). Small shallow flute marks (Fig. 6C), shallow narrow scours, and local tangential cross–beds (Fig. 6E) identify south-directed transport. We distinguish a separate arenite facies (*W3*, Table 1) that may represent the thicker (< 40 cm) sandstone beds of Quinn et al. (1999). It forms erosion-based beds, 20-40 cm in thickness, typically massive, some displaying normal grading. These deposits often occur capping coarsening-upward *W1-W2* bedsets in the upper 100 m (Fig. 5), and contain clasts of greenish siltstone (facies *W1*; Fig. 6D) that display soft-sediment deformation fabric.

We recognize three other bedform architectures in addition to planar beds. First, a 1-m-thick interval of convolute and syndepositionally faulted bedding bound by undisturbed stratigraphy at ~13 m above the formation base (Fig. 5). Second, from 45 to 60 metres above the formation base, strike-length tapering of sandstone and low–angle cross–lamination extends over several
metres (Fig. 6E). In this same interval is a paleochannel several metres in width with decimeter
paleorelief, filled with planar laminated siltstone (Fig. 6F). Another, much wider (30-50 m)
paleochannel, exposed at very low tide, and is filled with limestone conglomerate (see Fig. 12A).
The larger channels an approximate NW-SE bidirectional orientation (Fig. 5 inset).

**Environmental Interpretation**

The fine-scale alternation of siltstone and sandstone illustrates not just fluctuating energy levels
of transport and deposition, but an association between sand input and its transport under high-
energy (upper-flow) conditions as indicated by erosion-based sandstone beds with planar
lamination, current lineations, and flute marks (Collinson and Thompson, 1982). Abundant
evidence of vertical top-down bioturbation (Reynolds, 2015) illustrates restriction of seafloor
(re)colonization to periods of no or reduced sand accumulation (Chamberlain, 1971) that
accumulation as a product of some depositional event. The presence of bedding-top current
ripples and hummocky cross-stratification capping planar laminated sandstone was, in
combination with trace and body fossil material, interpreted by Quinn et al. (1999) as evidence
of storm deposition along a open shelf (Myrow and Southard, 1996).

Hummocky cross-stratification identifies a paleodepth near storm-wave base (Dott and
Bourgeois, 1982; Dumas and Arnott, 2006), but absolute depths are dependent on climatic
conditions (Peters and Loss, 2012). Local occurrence of interference ripples may indicate
shallower water than storm base, and changing water depths may be supported by the upsection
fining/thinning and coarsening/thickening bedsets that defines some longer-term change in a
combination of water depth, transport energy, and-or sediment influx.
Short-distance strike-length tapering of both individual sandstone beds and siltstone-sandstone
bedsets, and the presence of a large paleochannel oriented at a high-angle (NW-SE) to the
predominant southerly transport direction suggests that other sedimentary processes in addition
to storm transport were important. Whether the channels identify erosion related to sea-level fall
is unknown but the stratigraphic juxtaposition of channels and local tapered bedding may
illustrate development of small-scale downslope channel networks and spill-over (levee) deposits.
This type of transport network is best known related to deep sea fans (Pickering et al., 2008;
Arnott, 2010), but may also form where there is an abundant influx of sediment, such as along the
subaqueous extension of coastal fan-deltas (facies V and VI of Westcott and Ethridge, 1983), and
at the terminus of littoral sand-transport cells (Normandeau et al., 2013). In support of down-
gradient transport is the presence of intraclast-bearing facies W3, which is texturally similar to
debris and hyperconcentrated density flow deposits (Mulder and Alexander, 2001). This facies
incorporates fragments of the underlying compacted siltstone, and may represent flow in
response to instability along a bathymetric gradient. Numerous controls (e.g., oceanographic,
tectonic, depositional loading) can promote such movement (Shanmugam, 2016). Evidence for
structural instability in the Winterhouse succession is provided evidence of synsedimentary
deformation low in the section (Fig. 5), and the role of depositional loading might be indicated
by W3 beds capping coarsening-upward W1-W2 bedsets (Fig. 5) related to shallowing and-or
increased sediment input. We suggest, therefore, that this ancient shelf possessed a steeper
gradient than modern shelves (Myrow and Southard, 1996) enabling periods of offshore
transport. This may have occurred during storms or in response to changing water depth.
Lower Winterhouse (Carbonate, Mixed Sediment Succession)

Quinn et al. (1999) described attributes of two visually prominent facies interbedded with siliciclastic beds in the lower Winterhouse Formation: limestone conglomerate and calcarenite. We present some differences in description of these, along with new facies types, and their interpretation. Eight facies are grouped according to general texture (Table 3).

Calcisiltite

Fine-grained silt-size (< 20 microns) detrital calcite (facies $W4$) forms rare thin (mm-scale) deposits mantling other carbonate facies (see below), but also forms the matrix of a single bed (~20 cm thick) of skeletal rudstone ~10 m above the base of the formation (Fig. 7A, B). In this latter deposit, molds of gastropod and an indeterminate form are filled with calcite cement (Fig. 7A), and other skeletal grains are similar to those in facies $BP3c$ of the Lourdes Formation.

**Interpretation.** The detrital nature of the calcisiltite, as in unit BP3, requires transport of fine-grained carbonate from a source beyond the outcrop plane. Its rare presence might be related to elevated ambient bottom-water wave or current activity that precluded significant accumulation.

Skeletal and Sandy Skeletal Grainstones

Relatively pure skeletal-rich grainstone (facies $W5$) and sandy skeletal limestone (Mount, 1985) or sandy grainstone (facies $W6$) comprise sand- to granule-size deposits (Table 3) restricted to the lower ~100 m of the formation. Sandy grainstones are subdivided into two subfacies (Table 3) in which siliciclastics are either admixed (facies $W6a$) or finely interlaminated (facies $W6b$) with skeletal material. Facies $W5$ and $W6a$ form laterally discontinuous beds to lenses (< 1 m long) of massive, laminated, or normally graded fabrics.
Some normally graded beds preserve a thin (mm-scale) burrowed mantle of calcisiltite (facies \textit{W4}). Grainstone facies form single beds abruptly appearing in the siliciclastic succession, but more commonly are part of a reciprocal grainstone-siliciclastic bedset (Fig. 7C), and also underlie limestone breccia (facies \textit{W7}; Fig. 7D). Erosional bases can display cm-scale relief on siltstone, and one bed preserves a synsedimentary fracture filled by bounding siltstone (Fig. 7E). Facies \textit{W6b} differs lithologically from other deposits by displaying interlaminated siliciclastics and skeletal carbonate with a normal graded contact with overlying siltstone (Fig. 7F).

We differ from Quinn et al.'s (1999) assessment of skeletal preservation in grainstones. Thin sections show that skeletal (echinoderm, brachiopod, bivalve, trilobite) remains are strongly abraded and differentially blackened due to pyrite (Fig. 8A). There are two types of bivalve shells: locally blackened (pyritic) bivalve valves composed of crystalline mosaics of equigranular weakly ferroan calcite (Fig. 8A), with local ghost fibrous fabric visible under cross-polarized light; and, less numerous valves of non-ferroan prismatic calcite (Fig. 8B). The former identifies alteration of an original aragonitic bivalve shell (Flügel, 2010). The latter may represent external prismatic layers of an epifaunal pteriomorph (calcitic) bivalve (Esteban-Delgado et al., 2008).

Phosphatized skeletal material occurs throughout grainstone facies mostly associated with brownish discolouration of skeletal-grain margins (Fig. 8B) and along intra-skeletal (crinoid) microporosity. Scanning electron analysis reveals that the brownish colour is a combination of Fe-oxide and clusters to chains of apparent crystals and bodies of phosphate that emit a weak purplish-blue cathodoluminescence (Fig. 8C, D). Phosphate alteration predates all stages of calcite cement. Other occurrences of phosphatization include rare complete skeletal replacement, angular to rounded (bean-shaped) grains or pellets, and rare cement filling bryozoan zooecia.
Interpretation. Abraded and pyritic skeletal allochems, along with lime-mudstone lithoclasts, represent transport of material from a source not exposed in the Winterhouse outcrop belt. Rare mantles of calcisiltite identify some involvement of fine-fraction settling. Transport of blackened calcitic bivalve shells identify alteration at source, possibly through a similar shallow-burial/exhumation process interpreted for lithoclasts in the upper Lourdes Formation (see above). Variation between pure (facies $W_5$) and sandy (facies $W_6a$) compositions reveals either an erosional incorporation of siliciclastics during transport or transport of original mixed sediment, whereas facies $W_6b$ defines a high-order alternation of carbonate and siliciclastic deposition during waning transport energy. Quinn et al. (1999) interpreted a tempestite origin for the calcarenite deposits. However, hummocky cross-stratification or other wave-dominated fabrics characteristic of storm waning flow are not associated with the carbonates (Dott and Bourgeois, 1982). Normal grading does identify waning flow, but related to carbonate density flows (Mulder and Alexander, 2001; Playton et al., 2010) possibly arising from downslope transport triggered by storm deposition along an inner shelf (Dott and Bourgeois, 1982).

Limestone breccia

There are two limestone-breccia facies, $W_7$ and $W_8$ (Table 3). Facies $W_7$ is the most prominent, restricted to between ~20 m and ~100 m above the formation base. It forms metre-scale tabular bodies in dip view with individual breccia layers of < 1.5 m thickness. Initial appearance suggest strike-length continuity of beds; however, close examination shows that most beds consist of overlapping or laterally equivalent lens-like deposits tens of metres in width, laterally equivalent to siliciclastic successions (Fig. 9A; see also Quinn et al., 1999). These lenoid deposits also illustrate an upsection change in depositional loci (Fig. 9A). Breccia lenses
have NW to W axial orientations. Within a lens, breccia occurs interbedded with siltstone and
grainstone beds and lenses (Fig. 9B). Breccia deposits are abruptly overlain by siliciclastics,
including graptolitic shale (Fig. 9C; Bergström et al., 1974). Typically there is no downward
percolation of siliciclastic material but, in places, both clast and matrix paleoporosity display
geopetal clay drapes (Fig. 10A). Breccia clasts are similar in texture and composition to facies
W5 and W6a, as is the breccia matrix that typically exhibits much less intergranular
paleoporosity, including fully stylolitized contacts (Fig. 10B).

Breccia-clast thickness is < 40 cm. Larger clasts are tabular in shape (Fig. 9D), with long-axes
up to 150 cm. They are usually oriented subparallel to bedding; locally project above the
depositional bed plane; some appear rafted on underlying breccia (Fig. 9C); and, there is local
gentle (5-10°) westerly dipping imbrication developed (Fig. 9E; see also Shaikh, 1970;
Weerasingh, 1970; Quinn et al., 1999). Smaller clasts, ranging down to cobble size, tend to be
more rounded and are densely packed, often steeply oriented, between larger clasts.

In contrast to the above, facies W8 is represented by a single bed of skeletal grainstone (facies
W5) forming an in–plane breccia (Fig. 10C). Along the bedding plane is a transition from a fitted
block fabric into an area of greater interblock spacing, block imbrication, and out–of–plane block
rotation within siltstone (Fig. 10C). Interblock siltstone contains soft-sediment deformation
features, and is contiguous with an underlying bed. Siltstone disconformably overlies the breccia
(Fig. 10C).

**Interpretation.** Facies W7 is included in Quinn et al.’s (1999) *limestone conglomerate facies*
attributed by these authors to shoreface storm deposition in comparison with shelf flat-pebble
conglomerates (Markello and Read, 1981). The presence of strike-length lenses, upsection
change in depositional loci, absence of any significant waning flow texture capping these
deposits, and a markedly uncompacted grain fabric in clasts, but not matrix, suggest that their origin, as illustrated by facies W8, may have been associated with *autobrecciation* and mass transport down a gradient. The texture of facies W7 is similar to clast- to locally matrix-dominated debrites (Mulder and Alexander, 2001; Playton et al., 2010). Features such as rafted blocks and blocks projecting above the depositional plane occur in response to down-slope flow (e.g., Cook and Mullins, 1983). Locally exposed synsedimentary fractures in facies W5 (Fig. 7E), and development of breccia facies *W8*, demonstrates seafloor fracturing and movement of early-cemented limestone. The transition from fitted block fabric through increased inter-block spacing to out-of-plane block rotation is similar to developing breccia along the upslope limit of slide scars (Field and Clarke, 1979). We suggest that there is a genetic relationship between facies *W8* grading into facies *W7*, the latter occurring with increased abundance of transported skeletal carbonate to act as eventual matrix. In a thick unconsolidated skeletal sand, early cementation, differential movement, and downslope incorporation of blocks into a moving mass of skeletal sand will produce facies W7. The trigger for initiating debris flows remains uncertain but storm transport along an inner shelf could initiate localized seaward transport that initiates gravity flows down onto a deeper shelf substrate (Markello and Read, 1981). Imbricated blocks in facies *W8* spatially associated with syndeformation of siltstone support an equivalent origin for imbrication in facies *W7*, and as found in other debris flows (Major, 1998; Ilstad et al., 2004).

*Limestone conglomerate*

Conglomerate beds are usually < 30 cm thick, with an erosional base, and display massive to weakly graded and locally cross–bedded fabric. They are laterally confined over a strike width of 10s of metres. At ~50 metres above the formation base, a succession of conglomerate beds fill
and bevel a paleochannel cut into planar stratified siliciclastics with a paleorelief of 1-2 m (Fig. 11A, B). The facies is a monomictic (lime–mudstone) pebble to cobble matrix-support conglomerate (facies W9; Table 3; Fig. 11C upper), the matrix being a sandy skeletal grainstone similar to facies W6a. Lithoclasts are angular to rounded (Fig. 11C upper) consisting of a sandy micritic to microsparite texture (Fig. 11C lower) with abraded fossils (trilobites, crinoids). At ~180 metres above the formation base, facies W8 forms thin beds with erosional waveform surfaces of metre-scale wavelength that truncate south–directed cross–bedding (Fig. 11D).

**Interpretation.** The bimodal texture illustrates entrainment of two sediment sources. The massive to weakly graded texture, local south-directed cross-bedding, and laterally confined beds suggests that the conglomerate facies formed as a sandy debris flow (Shanmugan, 2012; Playton et al., 2010) or hyperconcentrated density flow (Mulder and Alexander, 2001). These flows would have followed local bathymetric lows as illustrated by conglomerate filling a paleochannel (Fig. 11A) oriented at a high angle to the predominant south-directed current. Similar deposits high in the section, however, illustrate southerly transport and may document deflection of a westerly flow by the predominant south-directed transport across the shelf.

**Shell-rich Concentrations**

There are two types. The first (facies W10) occurs only above 180 metres in the section, and consists of thin to moderately thick (< 30 cm) south-directed shingling of lenses to discontinuous beds of cement-rich skeletal (brachiopod, crinoid) and lithoclast-bearing cross-bedded grainstone and rudstone. These deposits have erosional bases and weakly defined south-directed cross-bedding (Fig. 12A, B). Accumulations of disarticulated brachiopod shells occur with shells overturned.
The second type (facies W11) includes locally exposed bedding-plane concentrations of skeletal material (Table 3) below ~20 metres and at ~180 metres above the formation base. In the lower 20 metres, small-area exposures reveal monospecific accumulations of brachiopods (whole and disarticulated), crinoid debris, and a layer of receptaculitids. At ~180 metres, a thin (cm-scale) concentration of skeletal (brachiopod–crinoid) material is admixed with very dark petrolierous siltstone.

**Interpretation.** Facies \textit{W10} illustrates a combination of \textit{in situ} accumulation of brachiopod clusters with net south-directed accumulation of mostly carbonate sediment interbedded with siliciclastics. Facies \textit{W11} represents autochthonous bedding-plane shell accumulations. Both types demonstrate evidence of wave and-or current influence.

### Carbonate Cements

**Types**

In unit BP3 (Lourdes Formation) and the lower Winterhouse Formation, intergranular paleoporosity is proportional to the amount of intergranular cement (< 40%) thereby illustrating the extent of mechanical compaction prior to lithification. Markedly uncompact grain texture occurs in facies \textit{W5}, clasts of this facies in limestone breccia (facies \textit{W7}), and related to brachiopod clusters in facies \textit{W10}. This paleoporosity is filled by a succession of six calcite cements, C1–C6 (Table 4). The sequence C1–C4 occupies the majority of intergranular paleoporosity, and is absent in siliciclastics of the Winterhouse Formation where, instead, type C5 and C6 cements are in common with later cementation in carbonate facies (Table 4).

C1 cement (Table 4) is represented by syntaxial calcite overgrowths. C1a is abundant, often occupying most available paleoporosity, and forms weakly ferroan equigranular overgrowths on
crinoid ossicles (Fig. 13A). CL is most often dark (see Fig. 8C and 13A), but dull to brighter CL defines local inter– and intracrystalline variation (Fig. 14A), especially where this cement lies adjacent to later C6 cement that has bright CL (Fig. 13B, E). This suggests that C1a cement has been differentially altered from an original composition. In contrast, C1b cement occurs only on calcitic bivalve shells (Fig. 9B), forming weakly ferroan prismatic crystals with dark CL.

C2 cement (Table 4) consists of thin (< 50 micron) to thick (<150 micron) isopachous layers of locally fibrous, but mostly microcrystalline calcite with speckled (bright, dark) cathodoluminescence (Fig. 13B, C). The cement fills intraskeletal microporosity of crinoid ossicles (Fig. 13A, B), and locally occurs interlaminated with syntaxial C1a cement (Fig. 13B). C2 calcite contains sub–micron solid (framboidal pyrite) inclusions.

C3 cement (Table 4) is locally very abundant, and occurs as discontinuous rims and crystal splayes of coarsely crystalline bladed or columnar calcite (Fig. 13). A relic finely fibrous fabric, visible using CL, is preserved in the lower parts of some crystal splays (Fig. 14A). The calcite has dark to dull CL (Fig. 13) that grades into diffuse patchy (dark, light) crystalline mosaics along crystal growth axes (Fig. 13A). Epifluorescence microscopy reveals green cores and gradational thin yellowish rims (Fig. 13D). This same pattern is associated with C4-cement crystals forming polygonal (including hexagonal) mosaics inter-grown with and succeeding C3 cement (Fig. 13D). C4 cement has dark CL (Fig. 13C).

C5 cement is ferroan, and of local distribution. It is followed by a more pervasive distribution of Fe-poor (C6) calcite (Table 4) that has very bright CL (Fig. 13B, E), occupies all remaining paleoporosity including late stage local microfractures oriented parallel to bedding.
Geochemistry

A fairly homogenous elemental geochemistry characterizes skeletal and diagenetic constituents (Appendix C). Calcite cements are low–Mg calcite (0.26 to 0.34 Mg wt %), and both staining and microprobe analyses illustrate a weakly ferroan (< 0.5 wt %) composition of all cements and calcitized bivalve fragments.

Compositional variation among cements is better demonstrated using stable ($\delta^{13}$C, $\delta^{18}$O) isotopes. C3 and C4 cements were sampled from large brachiopod-shelter paleoporosity in facies W10 (Fig. 14A inset photo) and uncompacted (~40 % paleoporosity; e.g., Fig. 8A) in a grainstone lithoclast in breccia at ~ 54 m. Oxygen isotope values overlap skeletal (brachiopod) and marine cement calcite reference fields for the Turinian and lower Chatfieldian stages (Fig. 15A). Most carbon-isotope values are more negative, up to 2.5 ‰, than these reference fields.

$^{87}$Sr/$^{86}$Sr ratios for C3 cement (mean = 0.708110, $\sigma$ = 0.000017; see Appendix C) are similar to early Chatfieldian marine calcite (Edwards et al., 2015).

Other stable isotope (C, O) data represent two-component mixtures due to more compacted depositional textures that preclude avoidance of cement-grain mixtures during sampling. C1 cement, calcitic (brachiopod) and once aragonitic skeletal material (algae, bivalves), and calcisiltite (facies BP3b) from the Lourdes Formation form a $\delta^{18}$O-$\delta^{13}$C cluster overlapping the range for Caradocian brachiopod calcite (Fig. 14A). A combined C5-C6 cement association forms a negative isotopic trend leading away from the marine reference fields (Fig. 14A), and this cement mixture contains a Sr–isotope ratio (~0.7088; see Appendix C) more radiogenic than Late Ordovician seawater.
Discussion

Physical Expression, Age, and Controls of Platform Drowning

The member-based stratigraphy of the Lourdes Formation records deepening, in stages, of an initial warm, shallow-water platform to a storm-influenced mid-ramp setting (Fig. 15; Stait and Barnes, 1991; Batten Hender and Dix, 2008). It culminates in a metre-scale-thick storm-influenced condensed succession (unit BP3); an interval of net reduced sediment-accumulation rates as represented by a high-order pattern of deposition, mechanical reworking, bioturbation, and periods of seafloor cementation yielding hardgrounds (Gómez and Fernández–López, 1994). In context of the entire Lourdes platform succession, unit BP3 is the sedimentary record of a drowning succession: initial incipient drowning (Kendall and Schlager, 1981; Dominguez et al., 1988) characterizing stratigraphic condensation, then platform termination as expressed by an abrupt drowning unconformity (Fig. 4B) characterized by superposition of marine-transported siliciclastics (Schlager, 1989; Erlich et al., 1990) of the Winterhouse Formation. Despite uncertainties in biostratigraphic age control (Fig. 2B), and certainly enhanced by the extended period of time defined by the condensed succession, platform drowning occurred within the Turinian-Chatfieldian interval of the Late Ordovician.

Possible driving forces leading to incipient, then final drowning are recorded in the Lourdes facies succession. First, interpreted storm beds (calci-siltite, grainstone, and conglomerate facies) contain not only a mixture of locally derived lithoclasts reworked from hardgrounds but fragments of skeletal grains and other lithoclasts (dolomite, lithic arenite) that identify sources positioned beyond the outcrop belt. Abraded crinoids and fragments of cyanobacteria and dasycladalean algae identify a normal-marine setting within the photic zone, and serve to identify continued carbonate production contemporary with the condensed succession. In context of
upsection deepening, this arrangement may identify a stepped-back platform and drowning of once mid-ramp setting (Fig. 15), a process well documented in other basins where abrupt rise in relative sea-level triggers a lateral shift in facies (Mullins et al., 1991; Galewsky et al., 1996; Menier et al., 2014).

However, such relative rise in sea level forms only a first-order control, and other factors of potential influence may have a more immediate effect (see Erlich et al., 1990). First, cooling of waters with increased deepening may be recorded by the upsection loss of coated grains (Fig. 2A), otherwise well defined warm-water indicators (Flügel, 2010). Likewise, a change in community structure of the coral *L. childensis*, from coral-microbial reefs in the middle member (James and Cuffey, 1986; Batten Hender and Dix, 2006) to sparsely distributed isolated colonies in the Beach Point Member, may also illustrate a prior thermal restriction of stable reef environments to shallower warmer (~18°C) water (Achituv and Dubinsky, 1990). Second, in the otherwise nutrient–poor Ordovician seas (Martin, 1996), deepening corresponds with increased presence of background siliciclastics illustrating net reduction in ambient wave and current energy. This may have increased stress on benthic filter feeders through elevated turbidity, nutrient loading, and promotion of a proliferation of bioturbators and borers (Hallock and Schlager, 1986; Fabricius, 2005). With bathymetric transition, potential changes in current patterns and transport of siliciclastic fines may have also influenced carbonate productivity (John and Mutti, 2005; Betzler et al., 2013).

A third possible control is the role of episodic seafloor hypoxia. Although a shallow burial-exhumation process may explain differentially blackened (pyritic) clasts and hardgrounds (Baird et al., 1986; Taylor and Wilson, 2003), the impact of seafloor hypoxia initiated by incursion of an oxygen minimum zone (OMZ) along an outer platform setting (Helly and Lavin, 2004) may
also be of influence. Fe content within OMZs is substantially elevated (< 350 ppm), especially in bottom waters on a shelf (Klar et al., in press), and would be sufficient to drive pyritization on the seafloor or in the water column under anoxic conditions (e.g., Bak and Sawlowicz, 2000).

Re-ventilation of the seafloor, such as during storm events, could rework blackened clasts into a transported oxic facies. With subsequent seafloor cementation, this provides an explanation for the final state of the Lourdes platform, facies BP3f (Fig. 3B).

Incipient drowning does not predict the impending burial of the Lourdes platform beneath Winterhouse siliciclastics. Instead, the exotic lithoclasts within unit BP3 herald this impending event. Grains of detrital dolomite and dolomitic arenite, and lithic arenite with chloritized serpentine identify sources related to older successions. The underlying thick siliciclastic succession of the Middle Ordovician Goose Tickle Group offers a ready and abundant supply of orogen-derived clastic material, but not dolomite (Quinn, 1992, 1995). Thin sections from Batten Hender (2007) demonstrate textural similarity between detrital dolomite in unit BP3 and dolomite cementing sandstone of the lowermost Lourdes Formation and lowermost Beach Point Member (Fig. 2A). The presence of this detrital material suggests unroofing and erosion of the platform interior and hinterland coupled with seaward drowning of the Lourdes platform.

Re-Establishment of Carbonate Production Along a Siliciclastic Seaway

The Winterhouse outcrop belt offers a narrow (< 20 m) stratigraphic perspective perpendicular to the strike of a newly developed offshore shelf succession (Quinn et al., 1999). In present-day coordinates, the siliciclastic shelf lay along the western margin of the developing Taconic Orogen (Quinn et al., 1999), facing a regional NE-SW-oriented seaway extending along the
Laurentian margin (Long and Copper, 1987). From this setting, the predominant (NE-SW, and S-directed) transport indicators characterize high-energy transport along the shelf to the southwest.

Reciprocal (carbonate, siliciclastic) stratigraphy of the lower Winterhouse succession was previously interpreted to document storm reworking of carbonate material derived from erosion of a source similar to the Lourdes platform (Quinn et al., 1999). This fits with coupling of hinterland uplift and seaward transport of sediment characterizing demise of the Lourdes platform, but only associated with reworking of detrital dolomite, and lime-mudstone lithoclasts in limestone conglomerate (Fig. 12). There should be a much greater abundance of sand-size grainstone lithoclasts incorporated into the various carbonate facies if the Lourdes Formation was the source of carbonate. We present an alternate sedimentary framework that incorporates rejuvenation of carbonate production contemporary with Winterhouse siliciclastic deposition.

Transport from a new (post-Lourdes) carbonate source is represented by calcisiltite (facies W4) low in the Winterhouse section; NW-SE-orientation of limestone breccia lenses and a paleochannel filled by limestone conglomerate (facies W9) parallel to the subordinate westerly transport indicators of interbedded siliciclastics (Fig. 5). Channelization within the siliciclastics, and interpretation of a shelf gradient (see above) support resedimented carbonates having been transported from a west-facing source within an inner-shelf setting. By 180 metres above the formation base, however, south-directed transport of carbonate may identify redirected flow parallel to the regional transport orientation (Fig. 15); a change that could arise from basin-margin retreat to the east relative to the plane of the outcrop belt (Fig. 15).

In terms of depositional facies models for carbonate deposition along a gradient (Cook and Mullins, 1983; Mullins and Cook, 1986; Playton et al, 2010), composition and texture of the resedimented grainstone and matrix of other facies identifies a skeletal-rich source. It was a
normal marine environment as indicated by an abundance of echinoderm debris that form
common and characteristic sediment contributors upon Late Ordovician platforms (Kiesling et al.,
2003). The source was subject to high energy as suggested by three indicators: (1) abundance of
abraded grains; the presence of (2) partially blackened (pyritized) grains; and (3) abraded Fe-
oxidized and phosphatized skeletal material that supports relict sediment and evidence of cycles
of sediment burial and exhumation (Cook and Shergold, 1986; James et al., 2004). The sparse
reciprocal stratigraphy and isolated lense-like breccia deposits may identify a depositional site
distal from source (Cook and Mullins, 1983; Mullins and Cook, 1986; Sheehan et al., 1993;
Playton et al., 2010). It remains uncertain if the carbonate source was a distinct paleoplatform
physically isolated from the southwesterly directed siliciclastic transport, or carbonate
production occurred along the inner shelf developed at times in response to sequestering of
siliciclastics during sea level highstands. In a tectonically active basin, there is an expected
heterogeneous distribution of along-strike and downslope architecture of siliciclastic, carbonate,
and mixed sediment depositional systems, each responding differently to changes in
accommodation space (Yose et al. 1989; Loucks and Handford, 1995; Bourget et al., 2013).
Limestone breccia (facies \W7\) may define downslope transport promoted during sea level
highstands when compared to mixing of rock clasts in limestone conglomerate (e.g., Loucks et al.
2011). However, shelf-gradient instability during highstands might also generate reworking and
mixing of all source types (Yose et al., 1989).

**Cool-Water Deposition and Cementation**

The common presence of partially phosphatized skeletal material and absence of any warm-
water indicators in the Winterhouse resedimented carbonate facies suggests that the source was
bathed by cool water by analogy with relatively contemporary sediments along the distal
Laurentian margin (e.g., Lavoie, 1995; Holland and Patzkowsky, 1996; Pope and Steffen, 2006).
Once deposited, rapid lithification occurred along the seafloor and in the near surface as evident
from an abrupt siliciclastic-carbonate contact with only rare downward sediment percolation, and
an uncompacted sediment fabric characterizing little mechanical compaction prior to lithification.

The C1-C4 cement succession is similar to marine or marine-derived cements from modern
and ancient platform settings (Table 4), including Ordovician hardgrounds (Wilkinson et al.,
1982; Marshall and Ashton, 1991) and cool-water deposits (Caron and Nelson, 2006). However,
the weakly ferroan composition and intracrystalline textural and CL variation suggests formation
and-or alteration under anoxic conditions such as in the near subsurface. CL variation within
syntactical (C1a) calcite cement crystals adjacent to brightly luminescent C6 cement characterizes
local diagenetic exchange of Mg by Mn, a principal activator of luminescence (Machel, 1985);
and the fibrous fabric of C2 and C3 cements are similar to modern marine and relict marine-
derived cementation (Schroeder, 1972; Marshall and Ashton, 1990; Major and Wilber, 1991;
Frank and Lohmann, 1996; Nelson and James, 2000; Caron and Nelson, 2006)

Despite alteration, the largely non-luminescent C3-C4 cement assemblage displays δ18O
values consistent with skeletal and marine cement reference fields for Turinian and Chatfieldian
limestone, an association supported by available biostratigraphic age control (Fig. 2B). The
majority of values for these cements lie between -4 and -3.2 ‰ (Fig. 14A). Using a range in
δ18Oseawater (−2 ‰ to −3 ‰) for this time period (see Tobin et al., 1996; Shields et al., 2003), the
estimated maximum range in isotope-based temperatures uncorrected for MgCO3 content is ~ 16
° to ~ 23° (Fig. 14B). Increased initial MgCO3 content in abiotic calcite results in an increase in
δ18O values (~0.26 ‰ per mol ‰; Tarutani et al., 1969) resulting in lower estimated temperatures
(Fig. 14B). These estimates are at least 7°C lower than sea-surface temperature estimates for the Turinian-Chatfieldian interval based on $\delta^{18}O_{apatite}$ (Trotter et al., 2008). This difference represents either deeper cooler water or identifies regionality in oceanographic conditions.

In this context, C3-C4 cements more depleted in $^{13}C$ than the Turinian-Chatfieldian reference fields (Fig. 15A) might be the signature of oxidation of organic matter with increased water depth along the shelf (Patterson and Walter, 1994). However, oxidation of organic carbon within near-surface pore-water can also generate a similar negative shift in isotope values (McCorkle et al., 1985; MacArthur et al., 1992).

Stable isotope and textural data support rapid near-surface cementation of transported carbonate along an interpreted shelf gradient at temperatures not unlike those associated with seasonal oceanic thermoclines in the subtropics (Turner and Kraus, 1967), the general paleolatitude of the study area in the Late Ordovician (Torsvik and Cocks, 2016). Subsequent C5–C6 cement stratigraphy in both Winterhouse siliciclastic and carbonate strata identifies a subsequent common burial diagenetic history. The occurrence of luminescent non-ferroan C6 cement in horizontal microfractures suggests a telogenetic stage of diagenesis (Choquette and Pray, 1970) driven by uplift and vertical extension.

**Platform Drowning: A View Across the Taconic Seaway**

A first-order driving force for drowning of the Lourdes platform along the Taconic Orogen was hinterland uplift coupled with seaward subsidence. This established both platform step-back and transition to cool-water carbonate production; then, eventual siliciclastic smothering of the remaining potential for carbonate production by the Edenian when, as indicated by
biostratigraphic control (Fig. 2B), ~700+ m of the remaining Winterhouse succession rapidly accumulated as part of a shallowing-upward siliciclastic succession (Quinn et al., 1999).

Along the distal Laurentian foreland margin, similar records of platform drowning beneath prograding orogen-derived siliciclastics are registered in the Chatfieldian-Edenian period in the form of repeated carbonate-siliciclastic cycles (Joy et al., 2000), and local margin collapse (Jacobi and Mitchell, 2002; Brett and Baird, 2002) related to longer term regional diachronous subsidence of the foreland margin in response to westward migration of the Taconic structural front (Ettensohn, 1991; Bradley and Kidd, 1991; Lavoie, 1994). During the Chatfieldian, relative sea-level rise caused a regional oceanographic transition to cool-water carbonate production that modified biotic communities (e.g., Brookfield, 1988; Keith, 1988; Lavoie, 1995; Holland and Patzkowsky, 1996; Patzkowsky and Holland, 1999). Similar patterns from either side of the Taconic seaway suggest a commonality related to flexural tectonic control. Improved biostratigraphic or other age control for the Long Point Group will be required to develop confident cross-basin correlations of foreland stratigraphies, but this would greatly improve understanding of proximal versus distal patterns of subsidence and sediment supply in response to flexural tectonics (Catuneanu et al., 1999) and development of the Taconic foreland basin.

**Conclusions**

Successive deepening through the Lourdes platform succession culminated with incipient drowning of a mid-ramp setting, represented by a metre-scale-thick condensed interval punctuated by storm deposits carrying skeletal material from an inferred stepped-back high-energy normal-marine platform (subsequently eroded). Transported lithoclastic detritus demonstrates erosion and transport of older (Lourdes platform, foreland-basin siliciclastic) strata.
suggesting that seaward subsidence was coeval with platform-interior and hinterland uplift. A drowning unconformity caps the condensed interval registered by marine transport of abundant siliciclastics. This transformed the carboante platform into a siliciclastic shelf.

Re-establishment of carbonate production within a new mid-shelf siliciclastic is registered by reciprocal stratigraphy of carbonate, mixed sediment, and siliciclastic sediment. Westerly-directed gravity-flow deposits of rare mud, skeletal-rich sand- and coarser textured carbonate and mixed sediment occur interbedded with storm-influenced siliciclastic sandstone dominated by NE-SW-oriented transport indicators. Composition and texture of re-sedimented skeletal-rich carbonate suggests that inner-shelf cool-water carbonate production occurred either isolated from siliciclastic inundation or controlled by change in sea level related sequestering of siliciclastics. Transported carbonate was rapidly cemented in cool (<16-23°C) waters down-gradient from the source; controls on transport remain uncertain.

Available biostratigraphic control suggests that the history of platform drowning, warm-to-cool water carbonate transformation, and eventual burial of carbonates beneath orogen-derived siliciclastics proximal to the Taconic orogen was generally contemporary with similar structural and stratigraphic patterns along the distal Laurentian margin during the Turinian-Edenian period. Improved age control for the Long Point succession will be needed to provide more detailed proximal-distal foreland-basin correlations.

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Figure Captions

Fig. 1 Bedrock geology, major structural elements, and outcrop localities along northwestern Port au Port Peninsula. Major fault trends: PBF, Piccadilly Bay Fault; RHT, Round Head Thrust; and TCT, Tea Cove Thrust. Bedrock geology map is based on Waldron et al. (2003).

Fig. 2 (A) Lithostratigraphic succession, members, and informal units of the Lourdes Formation (Stait and Barnes, 1991; Batten Hender and Dix, 2008) showing position and lithology of the newly defined informal unit BP3 of the Beach Point Member (this study). See Table 1 for lithofacies details. (B) Biostratigraphic age control based on Stait and Barnes (1991), Gillespie (1998), and Quinn et al. (1999), with solid lines denoting best estimate and dashed lines denoting potential range. For the Lourdes Formation, biota groups are (a) conodonts, (b) nautiloids, and (c) other shelly fauna. For the Winterhouse, biota groups are (d) graptolites and (e) achritarchs; meterage refers to distance above base of the formation.

Fig. 3 Unit BP3. (A) Loc. 1a (see Fig. 1) at low tide showing the thinly stratified character and base of the unit (arrow). Horizontal field of view is ~10 m. (B) Lithoclastic breccia (a: facies BP3e) of the Lourdes Formation overlain by fine–grained siliciclastics (b) of the Winterhouse Formation. Pocket knife (arrow) lies at the formation contact. (C) Bioturbated centimetre-scale bedding fabric (facies BP3a-c) illustrating relict grainstone lenses (arrows). Pocket knife (bottom) for scale. (D) Elevated portion of an erosional paleosurface (white arrow) with burrows within lime mudstone, and calcareous siltstone
(black arrow) resting on lower paleotopographic parts of the disconformity. Pocket knife for scale. (E) Bedding-plane view of scattered rounded clasts of lime mudstone and skeletal grainstone in calcareous siltstone. Several of the clasts display borings (arrow). Pen knife for scale. (F) Irregular erosional surface capping a skeletal grainstone (left arrow) and pyrite encrusted clast (right arrow) that occurs in siltstone immediately overlying the erosional surface. Pen knife for scale.

Fig. 4  (A) Sandy grainstone with differentially blackened lithoclast (right), smaller more delicate floc-like bacterial clasts (arrow), and siliciclastic grains (white). Thin-section photomicrograph, plane light, scale bar = 750 microns. B) Longitudinal section of tight helicoid tubular filament of *Obruchevella*. Thin-section photomicrograph, plane light, scale bar = 25 microns. C) Coccoid cells or transverse section of aligned tubes (black arrow) forming an isopachous encrustation on a shell fragment. Thin-section photomicrograph, plane light, scale bar = 25 microns. D) Detrital, zoned (pink = Fe-poor; blue = ferroan) crystalline dolomite in mixed sediment. Yellow grains are stained potassium feldspar. Thin-section photomicrograph, plane light, scale bar = 100 microns. E) Calcisiltite (facies *BP3b*) showing the irregular rounded to subrounded texture of calcite grains (white arrow) compared to a siliciclastic grain of similar size (black arrow). Scale bar = 20 microns. F) Limestone conglomerate (facies *BP3d*) with possible imbricated elongate pebbles near base. Scale bar is 20 cm. G) Grainstone lithoclast in limestone breccia (facies *BP3f*) showing blackened margin and boring walls (black arrows) and non-blackened fracture surface (white arrow). Scale bar is 1 cm.
Lithostratigraphy and paleocurrent data, lower Winterhouse Formation. General
distribution of facies, sedimentary structures, and fossil material in a composite section
based on localities 1 and 2, representing the exposed lower ~200 m of the Winterhouse
Formation. Not illustrated at this scale is higher–order reciprocal stratigraphy within
carbonate intervals as illustrated elsewhere. Paleocurrent data incorporates number of
observations per 10° of arc; tabulation of data and its treatment are presented in the
Appendix. Bidirectional data (n=28) include current lineations (x), and axial orientation
of a large paleochannel (a) filled with limestone conglomerate (facies W9). Unidirectional
transport indicators include flute marks (b), larger scours (c) filled with carbonate
sediment, and dip-direction of shingled carbonate lenses (d) (facies W10).

Siliciclastic facies attributes. (A) Interbedded succession of siltstone (facies W1) and
sandstone (facies W2) beneath carbonate strata (arrow). Vertical section is ~ 2 metres.
(B) Interference wave ripples. Pen knife for scale. (C) Bedding–plane exposure of
shallow flute marks. Width of stick is ~10 cm. (D) Single thick bed of sandstone (facies
W3) with siltstone clasts (dark). Hammer for scale. (E) South–directed tanhjential
crossbeds disconformably overlain by a low–angle south–tapering unit (double arrow )
of laminated sandstone (facies W2). Pocket knife for scale. (F) Cross-sectional dip-view
of a paleochannel filled with siltstone (a) defined by a discontinuous thin sandstone
(downward arrow) that disconformably overlie a sandstone-siltstone bedset (double
arrow). The north margin (to the right) of the channel overlies abrupt thickening of an
underlying sandstone bed. Hammer for scale.
Fig. 7 Sand–size skeletal grainstone (facies \( W5 \)) and mixed–sediment (facies \( W6 \)) facies. (A) Skeletal rudstone with lime mudstone matrix (facies \( W4 \)) showing horizontal‐elongate paleomold (centre) with geopetal lime mud. Dasycladalean alga occurs in the lower left, and calcitized bivalve shell overlies the paleomold. Inset box shows location of Fig. 7B. Plane light photomicrograph; scale bar = 500 microns. (B) Magnified view of inset box in Fig. 7A illustrating irregular rounded grain shapes of detrital calcite that comprise the lime mudstone. Plane light photomicrograph; scale bar = 50 microns. (C) Abrupt interstratification fabric of recessive siliciclastic and more resistant carbonate beds. Hammer for scale. (D) Interstratified siliciclastic and grainstone and mixed sediment beds (double arrow) preceding thick bed of carbonate breccia (a: facies \( W7 \)). Hammer for scale. (E) Synsedimentary fracture (black arrow), filled with siltstone and carbonate grains contiguous with bounding beds, in skeletal grainstone that has accumulated across erosional paleotopography (lower arrow) on compacted fine–grained siltstone (facies \( W1 \)). Knife end for scale. (F) Mixed–sediment (facies \( W6b \)) of interlaminated white skeletal‐rich layers and grey siltstone (c–s) forming a net “muddying” upward succession into siltstone (s). Pocket knife for scale.

Fig. 8 (A) Uncompacted texture of a skeletal grainstone bed with abraded and fragmented shells of brachiopods (pink), and differentially blackened once aragonitic (bivalve) allochems now consisting of weakly ferroan calcite. Stained thin section, plane light photomicrograph, scale bar = 500 microns. (B) Calcitic bivalve (a) with prismatic calcite structure, brownish (Fe‐oxide and phosphatic) rimmed margin (upper arrow), and syntaxial overgrowth cement (lower arrow). Cross‐polarized light; scale bar = 250
microns. (C) Purplish–blue luminescence of mineralized phosphate associated with intraskeletal microporosity of crinoid ossicles. This mineralization predates calcite syntaxial overgrowth cement with dark CL (white arrow). Scale bar = 250 microns. (D) SEM (backscatter) image showing phosphatic chains (greyish white) of crystal-like and rounded bodies. Scale bar = 50 microns.

Fig. 9  Limestone breccia, facies W7. (A) Outcrop view (width is ~20 m) showing two spatially distinct deposits of limestone breccia (arrows) at different stratigraphic levels, with interbedded sandstone and siltstone (s) laterally equivalent to the upper breccia deposit. (B) Succession of debrite beds (a–c) with intervening siltstone containing small tabular clasts (white arrow) and mound shaped grainstone lens (black arrow). Hammer for scale. (C) Abrupt contact between siltstone (arrow) across a large rafted clast resting on the top of limestone breccia. Hammer for scale. (D) Large clasts differentially oriented rafted on top of smaller clasts that typically display steeper orientations. Hammer for scale. (E) Local imbricate (W– to NW–dipping) fabric of large clasts. Hammer for scale.

Fig. 10  (A) Grainstone clast in limestone breccia (facies W7) displaying geopetral clay and detrital calcisiltite (upper arrow) occupying intergranular paleoporosity, and differentially compacted interlaminated clay and calcisiltite (lower arrow) with grainstone. Unstained thin section photomicrograph, plane light, scale bar = 250 microns. B) Compacted depositional texture showing intergranular stylolitic contacts. Thin-section photomicrograph, plane light, scale bar = 250 microns. (C) Bedding–plane exposure of in–plane limestone breccia (facies W8). The bed is disconformably overlain
by siltstone (white arrow) and illustrates increased interblock spacing (left to right) from a fitted fabric moving down the present westerly dipping bedding-plane exposure into an area of mostly siltstone and and imbricated blocks (arrows) displaying out–of–plane rotation. Hammer for scale.

Fig. 11 Limestone conglomerate (facies W9). (A) Stratigraphic discordance of the local strike of conglomerate beds (a, b), illustrated by white line extensions, filling a wide shallow paleo-channel cut down across the strike (double arrow) of underlying planar stratified siliciclastic strata. Horizontal view is ~50 m. (B) Close-up view of the angular discordance between conglomerate (a) and siliciclastic strata (s) at the channel margin. Hammer for scale. (C) (upper) Matrix-support monomictic lithoclastic texture. Hammer for scale. (lower) Texture and composition of lithoclast displaying sandy and silty micrite to microsparite matrix. Thin-section photomicrograph, plane light, scale bar = 200 microns. (D) Erosional mound geometries (arrows) of two beds of limestone conglomerate at 180 m above the formation base, displaying a possible metre-scale wavelength separating the mounds. In the upper bed, the erosional surface truncates a vague south–directed dip defined by lithoclasts.

Fig. 12 (A) Laterally discontinuous bed or thin lens of facies W10 (a) with vague south-directed cross-bedding, defined by inclined lithoclast molds (arrow), overlies disconformably laminated sandstone (b). Hammer for scale. (B) South–directed low-angle shingling of thin lenses of facies W10. Red and yellow arrows define the downdip and updip termination of lens a and b, respectively, separated by a thin siltstone bed. The
composite shingled interval is overlain by siltstone.

Fig. 13 Cement petrography. (A) Paired photomicrographs (top, cross–polarized transmitted; bottom, cathode luminescence): Top: Intergranular C3 cement (a) fills paleoporosity between brachiopod shell (b) and crinoid ossicles (C₁, C₂); scale bar = 250 microns. Bottom: The lower part of the crystal splay of C3 cement (black arrow) displays both a local fibrous fabric and gradation from dark–to–light (a) CL toward the outer part of the cement splay. The cement overlies a bright fringe of C2 cement on a brachiopod shell (b). The double black arrow points to contrasting dark and light CL associated with adjacent C1a cement overgrowths. Scale bar = 50 microns. (B) CL image providing an overview of succession of most abundant cements overlying the dark brachiopod shell (b): isopachous fringe of dark/bright C2 cement (white arrow); dark elongate crystals of C3 cement; and bright CL of C6 cement. Syntaxial overgrowth cement overlies an adjacent crinoid ossicle (C) which as C2 cement within its intraskeletal microporosity. Scale bar = 450 microns. (C) Isopachous distribution of patchy (dark, bright) CL of C2 cement (black arrow) coating skeletal material and overlain by dark CL of C3 cement. Scale bar = 75 microns. (D) Epifluorescence photomicrograph of elongate C3 crystals (white arrow) overlying a bivalve shell (s), and overlain by polygonal-shaped crystal mosaic of C4 cement (black arrow). The arrows also point to dark green crystal cores (white arrow) and yellowish green rims (black arrow). Scale bar = 100 microns. (D) CL photomicrograph showing brachiopod shells (b) overlain by bright, thin fringe of C2 cement, mostly dark CL of C3/C4 cements (a), and remaining void space occupied by brightly luminescent C6 cement (p). Scale bar = 250 microns.
Fig. 14  Stable isotope ($\delta^{13}$C, $\delta^{18}$O) data and temperature derivation. (A) Values for C3 and C4 cements sampled from uncompacted brachiopod textures of facies W10 and a clast of limestone breccia (facies W7) at 54 metres above the formation base. Inset photomicrograph shows a stained thin-section (scale bar = 250 microns) of brachiopods (pink) and intergranular cement (mauve). Other data sources are illustrated: skeletal grains and calcite cements in the lower Winterhouse formation; and calcitized dasycladacean algae and associated micrite/microspar textures in unit BP3 (Beach Point Member, Lourdes Formation). The field for Caradocian brachiopod calcite (Veizer, 1999) is illustrated along with two other Upper Ordovician reference sets: marine calcite cement (boxes; Tobin et al., 1996; 2005) from the lower Turinian Holston Formation (box 1), and mid-Chatfieldian Kullsberg Formation (box 2); and, brachiopod calcite (from Ludvigson et al., 1996): the upper Turinian Platteville Formation (a), lower Chatfieldian Decorah Formation (c, Sprechts Ferry Member; b, Guttenberg Member), an interpreted best estimate for Rocklandian marine signature (d), and Kullsberg Formation (e). Circle diameters define precision. (B) Potential range in paleotemperature based on C3 and C4 cements from ~54 and 180 metres above the formation base. The grey box (and red lines) show the maximum temperature based on calcite uncorrected for original MgCO$_3$ content. The extended dashed box and blue lines illustrate cooler temperature estimates with increased (up to 6–8 mol%) MgCO$_3$ content. Control points for $\delta^{18}$O$_{\text{seawater}}$: -2 to -3 % for the Late Ordovician (Tobin et al., 1996; Shields et al., 2003).
Fig. 15 Summary of sedimentological and oceanographic changes associated with drowning of the Lourdes platform, interpreted step-back of the carbonate source and downslope carbonate transport within the lower Winterhouse succession. Lourdes member stratigraphy: SP, Shore Point; BD, Black Duck; BP, Beach Point.
Fig. 1

238x197mm (300 x 300 DPI)
A

Winterhouse Fm

BP3

open platform, storm-

influenced

BP2

subtidal

BP1

peritidal

2

subtidal

10 m

Lourdes Formation

Black Duck

BP4

BD

1-3

SP4

SP3

SP2

SP1

Humber Arm allochthon

BP3f lithoclastic breccia

BP3e lithoclastic conglomerate

BP3a skeletal grainstone

BP3 facies (a-d) association:
siltstone, silty lime mudstone,
skeletal packstone/grainstone,
hardgrounds/erosional surfaces

B

Global

series/stages

North America

graptolite biozones

Britain

stages

Ma

complanatus

manitoulinensis

pygmaeus

spiniferus

ruedemanni americanus

bicornis

Ashgill

Katian

Cincinnatian

Richmondian

Mays-villian

Edenian

Sandbian

Mohawkian

Chattian

K

Sh

Lourdes

(b)

(c)

(d)

SP3-fm top

180-185 m,
& Loc. 4a,b

450

455

Middle Ordovician (Chazyan)

Winterhouse

Lourdes

84 -
144 m

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(renewed transgression)
Fig. 3

181x202mm (300 x 300 DPI)
Fig. 4

178x213mm (300 x 300 DPI)
**Facies and Lithology**

- W1&W2  siltstone & f.gr.arenite
- W3  med. gr. arenite
- W4-9  limestone

**Interbed**

- siltstone
- shale
- sandstone
- limestone

**Clast Type**

- W5, 6a  Lourdes

- fining upward, coarsening upward
- convolute bedding with synsedimentary fault
- exhumed mound shape
- wave cross-lamination
- hummocky lamination
- low-angle cross-beds

- brachiopod  nautiloid
- bryozoan  graptolite
- colonial coral  trilobite

**Paleocurrent Indicators:**

- a, x  bidirectional
- b-d  unidirectional
Fig. 7

177x200mm (300 x 300 DPI)
Fig. 10

159x191mm (300 x 300 DPI)
Fig. 11

177x251mm (300 x 300 DPI)
Fig. 12

126x128mm (300 x 300 DPI)
Fig. 13

177x225mm (300 x 300 DPI)
<table>
<thead>
<tr>
<th>Label</th>
<th>Lithology</th>
<th>Texture/Bedding</th>
<th>Boundaries</th>
<th>Sedimentary Features</th>
<th>Fossil/Other Material</th>
</tr>
</thead>
<tbody>
<tr>
<td>BP3-a</td>
<td>calcareous siltstone</td>
<td>biomottled, laminated;</td>
<td>burrowed,</td>
<td>discontinuous beds;</td>
<td>none observed</td>
</tr>
<tr>
<td></td>
<td></td>
<td>thin, thin beds</td>
<td>erosive surface</td>
<td>erosive surface</td>
<td></td>
</tr>
<tr>
<td>-b</td>
<td>lime mudstone (calcisiltite)</td>
<td>biomottled, abrupt base;</td>
<td>eroding surfaces,</td>
<td>erosional surfaces,</td>
<td>fragmented skeletal* debris</td>
</tr>
<tr>
<td></td>
<td></td>
<td>thin</td>
<td>hardground surfaces</td>
<td></td>
<td>calcified bacteria</td>
</tr>
<tr>
<td>-c</td>
<td>skeletal and lithoclastic</td>
<td>laminated, biomottled;</td>
<td>sharp base,</td>
<td>top-down bioturbation;</td>
<td>fragmented skeletal* fossils; calcified</td>
</tr>
<tr>
<td>grainstone</td>
<td>beds, lenses &lt; 10 cm thick</td>
<td>burrowed top</td>
<td>discontinuous beds</td>
<td></td>
<td>bacteria; lithoclasts (pyritic fossils);</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>detrital dolomite; phosphatic skeletal</td>
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<td></td>
<td></td>
<td></td>
<td>grains (rare)</td>
</tr>
<tr>
<td>-d</td>
<td>hardgrounds</td>
<td>pyritic (local);</td>
<td>a) undulating burrows,</td>
<td>burrows,</td>
<td>nautiloids, rugose coral, bryozoans</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>b) irregular borings</td>
<td></td>
<td>brachiopods lay scattered on surfaces</td>
</tr>
<tr>
<td>-e</td>
<td>lithoclastic conglomerate</td>
<td>matrix- to framework</td>
<td>sharp</td>
<td>local south-directed imbrication</td>
<td>clasts: BPb,c, and rare lithic arenite;</td>
</tr>
<tr>
<td></td>
<td></td>
<td>support</td>
<td></td>
<td></td>
<td>matrix: silt- and sand-size siliciclastics,</td>
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<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>local detrital dolomite, skeletal material*</td>
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</tr>
<tr>
<td>-f</td>
<td>lithoclastic breccia</td>
<td>matrix-support</td>
<td>sharp</td>
<td>pyritized angular clast margins and borings</td>
<td>clasts: facies BP3b,c; matrix: skeletal</td>
</tr>
<tr>
<td></td>
<td>(facies BP3c)</td>
<td></td>
<td></td>
<td></td>
<td>grainstone*, with siliciclastics, and rare</td>
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<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>phosphatic skeletal and rounded</td>
</tr>
</tbody>
</table>

* abraded crinoid ossicles and indeterminate echinoderm material; trilobites; brachiopods; calcitized bivalves; in BP3c and -f (matrix), the presence of calcareous algae (*Vermiporella* sp.) and the microproblematica fossil, *Nuia*;
<table>
<thead>
<tr>
<th>Label</th>
<th>Lithology</th>
<th>Texture/Bedding</th>
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<th>Fossil/Other Material</th>
</tr>
</thead>
<tbody>
<tr>
<td>W1</td>
<td>calcareous siltstone</td>
<td>5-15 cm thick;</td>
<td>usually sharp; interstratified</td>
<td>W1 and W2 form coarsening/thickening and fining/thinning metre-scale bedsets often capped by facies W3; local channels (metre to ten's of metre wide, and &lt;1 m deep)</td>
<td>W1 and W2 fossils: brachiopods, trilobites crinoids, bryozoa</td>
</tr>
<tr>
<td></td>
<td></td>
<td>sheet-like; laminated,</td>
<td>with W2</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>massive</td>
<td></td>
<td></td>
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</tr>
<tr>
<td>W2</td>
<td>sandstone (feldspathic-quartz arenite)</td>
<td>5-15 cm thick;</td>
<td>usually sharp; interstratified</td>
<td>W2 beds display current lineations, flute marks, HCS, wave/current ripple marks/lamination; local metre-scale tapereding; local channels</td>
<td>W1 and W2 burrows: vertical and horizontal, biomottled fabrics top-down bioturbation prevalent in upper 100 m on sandstone beds</td>
</tr>
<tr>
<td></td>
<td></td>
<td>laminated, massive; local</td>
<td>with W1</td>
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<td></td>
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<tr>
<td></td>
<td></td>
<td>lenses</td>
<td></td>
<td></td>
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</tr>
<tr>
<td>W3</td>
<td>sandstone (feldspathic-quartz arenite)</td>
<td>medium-grained; sharp</td>
<td>local individual to clustered</td>
<td>local individual to clustered siltstone (W1) clasts showing evidence of soft-sediment deformation</td>
<td>fossil fragments; local detrital dolomite</td>
</tr>
<tr>
<td></td>
<td></td>
<td>massive, rarely graded</td>
<td>siltstone (W1) clasts showing</td>
<td></td>
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<tr>
<td></td>
<td></td>
<td></td>
<td>evidence of soft-sediment</td>
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<td></td>
<td></td>
<td></td>
<td>deformation</td>
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Table 3. Carbonate and mixed sediment facies, lower Winterhouse Formation

<table>
<thead>
<tr>
<th>Label</th>
<th>Lithology</th>
<th>Texture/Bedding</th>
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<tbody>
<tr>
<td><strong>Calcisiltite</strong></td>
<td></td>
<td></td>
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</tr>
<tr>
<td>W4</td>
<td>skeletal rudstone</td>
<td>single bed (20 cm); massive</td>
<td>sharp</td>
<td>calcisiltite matrix</td>
<td>molds of gastropod and indeterminate form; fragments of brachiopods, trilobites, calcitized bivalves, dasycladacean algae</td>
</tr>
<tr>
<td><strong>Grainstone</strong></td>
<td></td>
<td></td>
<td></td>
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<td></td>
</tr>
<tr>
<td>W5</td>
<td>skeletal grainstone</td>
<td>10-20 cm thick beds; and lenses; massive, laminated, rarely graded</td>
<td>sharp</td>
<td>local bioturbated tops, rare mantles of calcisiltite; uncompacted texture common</td>
<td>fossil fragments: crinoids, calcitic and calcitized bivalves, trilobites, cryptostomid bryozoans; lithoclasts: rare BP3b clasts; phosphatized fossil fragments</td>
</tr>
<tr>
<td><strong>Mixed Sediment</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
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</tr>
<tr>
<td>W6</td>
<td>sandy skeletal grainstone</td>
<td>&lt; 30 cm beds; two subfacies: 6a, 6b</td>
<td>sharp base</td>
<td>local bioturbated tops, siliciclastics admixed with skeletal carbonate, siliciclastics finely interlaminated with skeletal carbonate, grades up into siltstone and calcisiltite</td>
<td>same as W5Avoid</td>
</tr>
<tr>
<td><strong>Limestone Breccia</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>W7</td>
<td>limestone breccia</td>
<td>massive; &lt;1.5 m thick; framework- to matrix-support, with skeletal grainstone matrix;</td>
<td>sharp; base is erosional</td>
<td>clast imbrication (westerly); rare geopetal clay and calcisiltite in paleoporosity in clasts and matrix; rafted</td>
<td>fossil fragments and clast lithology as in facies W5 and W6a; matrix as in facies W5</td>
</tr>
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<table>
<thead>
<tr>
<th>Facies</th>
<th>Description</th>
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<tbody>
<tr>
<td>W8</td>
<td>in-plane limestone breccia; single bed (15 cm thick; facies W5)</td>
</tr>
<tr>
<td>Limestone Conglomerate</td>
<td></td>
</tr>
<tr>
<td>W9</td>
<td>limestone conglomerate; thin (&lt; 30 cm) beds; massive to graded; sandy skeletal grainstone matrix; laterally restricted sheet deposits</td>
</tr>
<tr>
<td>Skeletal Concentrations</td>
<td></td>
</tr>
<tr>
<td>W10</td>
<td>skeletal rudstone; lenses (m-scale length; &lt; 30 cm thickness; sandy skeletal grainstone matrix)</td>
</tr>
<tr>
<td>W11</td>
<td>bedding-plane concentrations; millimetres to few centimetres in thickness</td>
</tr>
<tr>
<td>Cement Type</td>
<td>Form &amp; Size</td>
</tr>
<tr>
<td>-----------------</td>
<td>------------------------------</td>
</tr>
<tr>
<td><strong>Restricted to carbonate facies</strong></td>
<td></td>
</tr>
<tr>
<td><strong>C1 syntaxial overgrowth</strong></td>
<td></td>
</tr>
<tr>
<td>a: on crinoid ossicle</td>
<td>equigranular; &lt; 2000 L x W(^2)</td>
</tr>
<tr>
<td>b: on calcitic bivalve shell</td>
<td>palisade (fibrous) 50-100L; 10-20W</td>
</tr>
<tr>
<td><strong>C2 fibrous</strong></td>
<td>isopachous rims 35-50L; &lt; 10W</td>
</tr>
<tr>
<td><strong>C3 bladed</strong></td>
<td>individual splays; &lt; 600L; 50-200W</td>
</tr>
<tr>
<td><strong>C4 polygonal mosaic</strong></td>
<td>euhedral, mosaic &lt; 100L; 50-60W</td>
</tr>
<tr>
<td><strong>Common to carbonate and siliciclastics facies</strong></td>
<td></td>
</tr>
<tr>
<td><strong>C5 (ferroan)</strong></td>
<td>equigranular &lt; 1000 L x W</td>
</tr>
<tr>
<td><strong>C6 (non-ferroan)</strong></td>
<td>equigranular &lt; 1000 L x W</td>
</tr>
</tbody>
</table>

1 according to staining and Fe concentrations (Table 3); abbreviations: CL, cathodoluminescence; F, fluorescence.
2 L, length; W, width.
### Appendix A: Geographic coordinates of localities

<table>
<thead>
<tr>
<th>LOCATION</th>
<th>STRATIGRAPHIC REFERENCE</th>
<th>GPS COORDINATES</th>
<th>MAP SHEET*</th>
<th>REFERENCE**</th>
</tr>
</thead>
<tbody>
<tr>
<td>1A.</td>
<td>Lourdes-Winterhouse formation contact</td>
<td>48.7836°N, 58.7693°W</td>
<td>12B15</td>
<td>a</td>
</tr>
<tr>
<td>1B.</td>
<td>breccia lobes, lower Winterhouse Formation</td>
<td>48.7762°N, 58.7794°W</td>
<td>12B15</td>
<td>a</td>
</tr>
<tr>
<td>1C.</td>
<td>channel (offshore) lower Winterhouse Formation</td>
<td>48.7769°N, 58.7792°W</td>
<td>12B15</td>
<td>a</td>
</tr>
<tr>
<td>1D.</td>
<td>breccia, lower Winterhouse Formation</td>
<td>48.7757°N, 58.7804°W</td>
<td>12B15</td>
<td>a</td>
</tr>
<tr>
<td>Loc. 2</td>
<td>upper 40 m of the lower Winterhouse Formation</td>
<td>48.7198°N, 58.8695°W</td>
<td>12B10</td>
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</tr>
<tr>
<td>Loc. 3</td>
<td>Winterhouse-Misty Point formation contact§  (inverted section: Three Rock Point)</td>
<td>48.6268°N, 59.0988°W</td>
<td>12B11</td>
<td>§</td>
</tr>
<tr>
<td>Loc. 4</td>
<td>Winterhouse-Misty Point formation contact  (north end of Clam Bank Cove)</td>
<td>48.6544°N, 58.9986°W</td>
<td>12B11</td>
<td>a</td>
</tr>
</tbody>
</table>

* UTM Grid Zone 21, 1:250,000; ** a, part of the Shoreline section of Quinn et al. (1999); § exposed only at very low (Spring) tide
Appendix B: Paleocurrent Data

Stratigraphic position (metres) is above the base of the Winterhouse Formation. Multiple measurements for a given position identify different bedding planes. Linear data are not corrected for dip: an estimated correction (< 5°) for the maximum dip (34°) combined with 10°-bin grouping of data does not alter the raw paleogeographic bimodality nor unidirectional orientations (see Fig. 5). Correction for dip orientation of downlapping carbonate (skeletal-cement) lenses at 181 m followed Tucker (1988).

<table>
<thead>
<tr>
<th>Metres</th>
<th>Azimuth</th>
<th>Type</th>
<th>Facies</th>
<th>Bedding (Strike/Dip)</th>
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<tr>
<td>4</td>
<td>038-218</td>
<td>current lineation</td>
<td>W2</td>
<td>200/20 NW</td>
</tr>
<tr>
<td>30</td>
<td>160-340</td>
<td>wave ripple crest</td>
<td>W2</td>
<td>227/20 NW</td>
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<tr>
<td>39</td>
<td>062-242</td>
<td>current lineation</td>
<td>W2</td>
<td>220/15 NW</td>
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<tr>
<td>48</td>
<td>300</td>
<td>small paleoscour</td>
<td>W3</td>
<td>220/18 NW</td>
</tr>
<tr>
<td>55</td>
<td>135-315</td>
<td>current lineation</td>
<td>W2</td>
<td>220/22 NW</td>
</tr>
<tr>
<td>150-330</td>
<td>current lineation</td>
<td>W2</td>
<td>220/22 NW</td>
<td></td>
</tr>
<tr>
<td>68</td>
<td>240</td>
<td>trough cross-bed</td>
<td>W2</td>
<td>222/20 NW</td>
</tr>
<tr>
<td>70</td>
<td>130-310</td>
<td>current lineation</td>
<td>W2</td>
<td>222/20 NW</td>
</tr>
<tr>
<td>89</td>
<td>030-210</td>
<td>current lineation</td>
<td>W2</td>
<td>220/18 NW</td>
</tr>
<tr>
<td>050-230</td>
<td>current lineation</td>
<td>W2</td>
<td>220/18 NW</td>
<td></td>
</tr>
<tr>
<td>90</td>
<td>050-230</td>
<td>current lineation</td>
<td>W3</td>
<td>224/18 NW</td>
</tr>
<tr>
<td>91</td>
<td>240</td>
<td>current scour</td>
<td>W2</td>
<td>224/18 NW</td>
</tr>
<tr>
<td>92</td>
<td>050-230</td>
<td>current lineation</td>
<td>W2</td>
<td>224/18 NW</td>
</tr>
<tr>
<td>060-240</td>
<td>current lineation</td>
<td>W2</td>
<td>224/18 NW</td>
<td></td>
</tr>
<tr>
<td>055-235</td>
<td>current lineation</td>
<td>W2</td>
<td>224/18 NW</td>
<td></td>
</tr>
<tr>
<td>050-230</td>
<td>current lineation</td>
<td>W2</td>
<td>224/18 NW</td>
<td></td>
</tr>
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<td>060-240</td>
<td>current lineation</td>
<td>W2</td>
<td>224/18 NW</td>
<td></td>
</tr>
<tr>
<td>97</td>
<td>040-220</td>
<td>current lineation</td>
<td>W2</td>
<td>224/22 NW</td>
</tr>
<tr>
<td>150-330</td>
<td>current lineation</td>
<td>W2</td>
<td>224/22 NW</td>
<td></td>
</tr>
<tr>
<td>102</td>
<td>130-310</td>
<td>current lineation</td>
<td>W2</td>
<td>224/22 NW</td>
</tr>
<tr>
<td>107</td>
<td>010-190</td>
<td>current lineation</td>
<td>W2</td>
<td>224/20 NW</td>
</tr>
<tr>
<td>070-250</td>
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<td>W2</td>
<td>224/20 NW</td>
<td></td>
</tr>
<tr>
<td>118</td>
<td>240</td>
<td>flute/scour</td>
<td>W2</td>
<td>228/20 NW</td>
</tr>
<tr>
<td>120</td>
<td>230</td>
<td>flute/scour</td>
<td>W2</td>
<td>226/22 NW</td>
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<tr>
<td>134</td>
<td>170-350</td>
<td>current lineation</td>
<td>W3</td>
<td>222/30 NW</td>
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<tr>
<td>137</td>
<td>080-260</td>
<td>current lineation</td>
<td>W3</td>
<td>222/30 NW</td>
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<tr>
<td>145</td>
<td>080-260</td>
<td>current lineation</td>
<td>W2</td>
<td>220/32 NW</td>
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<td>220/32 NW</td>
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<tr>
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<td>W3</td>
<td>226/30 NW</td>
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<tr>
<td>164</td>
<td>130-310</td>
<td>current lineation</td>
<td>W2</td>
<td>226/30 NW</td>
</tr>
<tr>
<td>170</td>
<td>145-325</td>
<td>current lineation</td>
<td>W2</td>
<td>228/34 NW</td>
</tr>
<tr>
<td>181</td>
<td>180</td>
<td>dip of carbonate lens</td>
<td>W7</td>
<td>228/30 NW</td>
</tr>
<tr>
<td>180</td>
<td>dip of carbonate lens</td>
<td>W7</td>
<td>228/30 NW</td>
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000/15W
Appendix C: Geochemistry  
a. trace element concentrations versus stain colour

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<th></th>
<th>Mg</th>
<th>Mn</th>
<th>Fe</th>
<th>Sr</th>
<th>Ba</th>
<th>Na</th>
<th>Stain*</th>
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<td><strong>Skeletal, sediment types</strong></td>
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<td></td>
<td></td>
<td></td>
<td></td>
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</tr>
<tr>
<td>bivalve (arag.)</td>
<td>0.21</td>
<td>0.13</td>
<td>0.30</td>
<td>0.04</td>
<td>0.01</td>
<td>0.01</td>
<td>light purple (5PB7/6)</td>
</tr>
<tr>
<td>(n=6)</td>
<td>0.06</td>
<td>0.05</td>
<td>0.03</td>
<td>0.03</td>
<td>0.02</td>
<td>0.01</td>
<td></td>
</tr>
<tr>
<td>bivalve (calcitic)</td>
<td>0.91</td>
<td>0.06</td>
<td>0.11</td>
<td>0.25</td>
<td>0.02</td>
<td>0.03</td>
<td>pink (5R5/10)</td>
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<tr>
<td>(n=4)</td>
<td>0.41</td>
<td>0.08</td>
<td>0.08</td>
<td>0.14</td>
<td>0.02</td>
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<tr>
<td>brachiopod</td>
<td>0.34</td>
<td>0.03</td>
<td>0.07</td>
<td>0.12</td>
<td>0.01</td>
<td>0.07</td>
<td>pink</td>
</tr>
<tr>
<td>(n=11)</td>
<td>0.05</td>
<td>0.05</td>
<td>0.10</td>
<td>0.05</td>
<td>0.02</td>
<td>0.03</td>
<td></td>
</tr>
<tr>
<td>crinoid ossicle</td>
<td>0.29</td>
<td>0.09</td>
<td>0.29</td>
<td>0.04</td>
<td>0.00</td>
<td>0.01</td>
<td>pink, light purple</td>
</tr>
<tr>
<td>(n=6)</td>
<td>0.07</td>
<td>0.07</td>
<td>0.26</td>
<td>0.03</td>
<td>0.00</td>
<td>0.01</td>
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</tr>
<tr>
<td>dasycladaceans</td>
<td>0.38</td>
<td>0.05</td>
<td>0.26</td>
<td>0.05</td>
<td>0.01</td>
<td>0.00</td>
<td>light purple</td>
</tr>
<tr>
<td>(n=6)</td>
<td>0.13</td>
<td>0.02</td>
<td>0.06</td>
<td>0.04</td>
<td>0.01</td>
<td>0.00</td>
<td></td>
</tr>
<tr>
<td>micrite/microspar</td>
<td>0.42</td>
<td>0.06</td>
<td>0.19</td>
<td>0.02</td>
<td>0.02</td>
<td>0.01</td>
<td>light purple</td>
</tr>
<tr>
<td>(n=9)</td>
<td>0.10</td>
<td>0.02</td>
<td>0.03</td>
<td>0.01</td>
<td>0.02</td>
<td>0.01</td>
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</tr>
<tr>
<td><strong>Cement types</strong></td>
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<td></td>
<td></td>
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<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>C1a (n=21)</td>
<td>0.28</td>
<td>0.13</td>
<td>0.31</td>
<td>0.02</td>
<td>0.01</td>
<td>0.01</td>
<td>light purple</td>
</tr>
<tr>
<td></td>
<td>0.09</td>
<td>0.06</td>
<td>0.18</td>
<td>0.04</td>
<td>0.01</td>
<td>0.01</td>
<td></td>
</tr>
<tr>
<td>C2 (n=14)</td>
<td>0.29</td>
<td>0.15</td>
<td>0.39</td>
<td>0.02</td>
<td>0.01</td>
<td>0.02</td>
<td>light purple</td>
</tr>
<tr>
<td></td>
<td>0.12</td>
<td>0.04</td>
<td>0.21</td>
<td>0.03</td>
<td>0.01</td>
<td>0.02</td>
<td></td>
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<tr>
<td>C3 (n=22)</td>
<td>0.31</td>
<td>0.07</td>
<td>0.38</td>
<td>0.06</td>
<td>0.00</td>
<td>0.01</td>
<td>light purple</td>
</tr>
<tr>
<td></td>
<td>0.17</td>
<td>0.05</td>
<td>0.28</td>
<td>0.05</td>
<td>0.01</td>
<td>0.01</td>
<td></td>
</tr>
<tr>
<td>C4 (n=34)</td>
<td>0.34</td>
<td>0.10</td>
<td>0.40</td>
<td>0.06</td>
<td>0.01</td>
<td>0.01</td>
<td>light purple</td>
</tr>
<tr>
<td></td>
<td>0.11</td>
<td>0.06</td>
<td>0.14</td>
<td>0.06</td>
<td>0.01</td>
<td>0.00</td>
<td></td>
</tr>
<tr>
<td>C5 (n=35)</td>
<td>0.28</td>
<td>0.15</td>
<td>0.22</td>
<td>0.03</td>
<td>0.01</td>
<td>0.01</td>
<td>light to dark purple (5PB7/6 to 5PB4/6)</td>
</tr>
<tr>
<td></td>
<td>0.09</td>
<td>0.10</td>
<td>0.10</td>
<td>0.05</td>
<td>0.02</td>
<td>0.01</td>
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</tr>
<tr>
<td>C6 (n=7)</td>
<td>0.24</td>
<td>0.14</td>
<td>0.11</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>pink</td>
</tr>
<tr>
<td></td>
<td>0.13</td>
<td>0.10</td>
<td>0.02</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td></td>
</tr>
</tbody>
</table>

* verbal description, and equivalent Munsell colour number
b) isotope data*

<table>
<thead>
<tr>
<th>LOURDES FORMATION</th>
<th>C4 cement</th>
<th>C6 cement</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Micrite (m), microspar (mi)</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>KP-5 mi</td>
<td>-0.4</td>
<td>26.7</td>
</tr>
<tr>
<td>W8a mi</td>
<td>-1.1</td>
<td>26.1</td>
</tr>
<tr>
<td>W15 mi</td>
<td>-0.6</td>
<td>25.9</td>
</tr>
<tr>
<td>W22a mi</td>
<td>-1.2</td>
<td>25.6</td>
</tr>
<tr>
<td>W24 mi</td>
<td>-0.9</td>
<td>26.1</td>
</tr>
<tr>
<td>W25 mi</td>
<td>-1.5</td>
<td>26.2</td>
</tr>
<tr>
<td>W25a mi</td>
<td>-1</td>
<td>26.4</td>
</tr>
<tr>
<td>W25c mi</td>
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<td>25.8</td>
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<td>W29 m</td>
<td>-1.2</td>
<td>25.3</td>
</tr>
<tr>
<td>W53d m</td>
<td>-0.7</td>
<td>25.7</td>
</tr>
</tbody>
</table>

| Calcitized dasycladacean fragments | | |
| W22 | -0.9 | 26.1 | -4.67 |
| W25b | -1.5 | 25.3 | -5.44 |
| W53a | -1.1 | 25.4 | -5.34 |
| W53b | -1.1 | 25.4 | -5.34 |

<table>
<thead>
<tr>
<th>LOWER WINTERHOUSE FORMATION</th>
<th>Selective dissolution, dolomite + C6 cement</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Brachiopod</strong></td>
<td></td>
</tr>
<tr>
<td>W3</td>
<td>-0.5</td>
</tr>
<tr>
<td>KP-2</td>
<td>0.5</td>
</tr>
</tbody>
</table>

| Bivalve (calcitized, ferroan) | | |
| W32 | -0.8 | 25.6 | -5.15 |
| W18 | 0.1 | 24.8 | -5.93 |
| W10 | -1.0 | 25.4 | -5.34 |
| W6b | -0.8 | 25.5 | -5.25 |

| C1 Cement and crinoid mix | | |
| W2b | -0.8 | 25.4 | -5.34 |
| W5 | -1.1 | 25.6 | -5.15 |
| W17 | -0.7 | 25.7 | -5.05 |
| W31b | -0.6 | 25.6 | -5.15 |
| W42a | -1.2 | 25.1 | -5.64 |

| C3 cement | | |
| KP-3 | 0.4 | 27.1 | -3.70 |
| W31a | -0.2 | 27 | -3.79 |
| W33 | -0.2 | 27.5 | -3.31 |
| W50a | -0.1 | 27.2 | -3.60 |
| W50b | 0.0 | 27.1 | -3.70 |
| W50d | -0.1 | 27.3 | -3.50 |

| **87Sr/86Sr** | | |
| W50a | 0.70810066 | C3 cement |
| W50d | 0.70809934 | C3 cement |
| W50b | 0.70810546 | C4 cement |
| W50c | 0.70813480 | C4 cement |
| W31 | 0.70882487 | C5-C6 cement |

*std error: \( \delta^{13}C \), 0.01‰; \( \delta^{18}O \), 0.01‰; \( \frac{87}{86}\)Sr, 0.00000009 to 0.00000012