Basement influences on dolomite-hosted vertical sedimentary intrusions in marine erg-margin deposits from the Pennsylvanian of northern Wyoming (USA)

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ABSTRACT

Understanding soft-sediment deformation structures and their triggers can help in assessing the influence of tectonics, climate, and diagenesis on the stratigraphic record. Such features commonly record processes that would not otherwise be preserved. The description of soft-sediment deformation in Pennsylvanian deposits of the western United States, characterized by orbitally-driven alternations between eolian sandstones, marine dolomites and shales, has been limited to contorted cross-beds. We document discordant, sheet-like sedimentary intrusions in three marine intervals over a 45 km-wide area. Intrusions consist of very well to moderately cemented, very-fine to fine-grained quartz sandstone. Body widths range from 5 to 50 cm, and heights up to 2 m. The orientations of 103 vertical bodies were measured. Based on upward- and downward-tapering, and the presence of deformed, microfractured fragments of host rocks, these intrusions are interpreted to result from seismically-induced fluidization of water-saturated sands. Their sheet-like morphology indicates injection through fractures. Two predominant directions (WNW-ESE and N-S) were recognized and interpreted as pre-injection fracture sets. Folding of surrounding layers around the intrusions suggests negligible compaction prior to injection, indicating penecontemporaneous or shallow burial fluidization. The intraplate location of Wyoming implies that seismicity did not originate at a plate boundary. The area within which intrusions are found is crossed by a zone characterized by localized development of thick eolian stories at the top of the formation,

interpreted to reflect the rejuvenation of a basement lineament. The seismically-active character of lineaments may explain overpressure and fluidization, substantiating the notion that Precambrian structures repeatedly affected Phanerozoic sedimentation. Similar intrusive features may be wrongly identified or overlooked in deposits of arid environments, but their recognition provides insights into structural context during deposition and burial.

Keywords: Injectites; Planar sandstone dikes; Seismically-induced fluidization; Eolian sandstones; Basement lineament

1. INTRODUCTION

Following a long history of research on the link between soft-sediment deformation and its triggers (e.g. Seilacher, 1969; Lowe, 1975; Doe and Dott, 1980; Owen, 1996; Montenat et al., 2007; Owen and Moretti, 2011; Owen et al., 2011), interest in intrusive sandstone bodies peaked in the 1990s and 2000s after their relevance to hydrocarbon exploration was recognized (Hurst and Cartwright, 2007). Vertical sedimentary intrusions are known from a wide range of environments: subglacial (Rijsdijk et al., 1999; Le Heron and Etienne, 2005; Ravier et al., 2015a), deep marine (Strachan, 2002; Duranti and Hurst, 2004; Parize et al., 2007), fluvio-lacustrine (Aspler and Donaldson, 1986; Li et al., 1996; Rodríguez-Pascua et al., 2000), and desert ergs (Huuse et al., 2005; Chan et al., 2007; Ross et al., 2014). Soft-sediment deformation structures are especially common-place in eolian depositional environments (e.g. Doe and Dott, 1980; Horowitz, 1982; Loope et al., 2001; Netoff, 2002; Chan et al., 2007; Loope et al., 2013), due to the combination of loose packing arrangement of grains during deposition and variations in the water table. In such environments, soft-sediment deformation structures include contorted bedding and laminations (Doe and Dott, 1980; Horowitz, 1982), vertical sandstone dikes (Bryant and Miall, 2010),

and large (up to 100 m diameter) columnar intrusions (Netoff and Shroba, 2001; Netoff, 2002; Chan et al., 2007).

Caused by the movement of fluidized sand, these features record periods of anomalously high pore-water pressure in water-saturated sand (Jolly and Lonergan, 2002; Hurst et al., 2011). A wide array of processes has been linked to overpressure and soft-sediment deformation, ranging from penecontemporaneous events such as storms, waves, tides, and ice loading (e.g. Dalrymple, 1979; Martel and Gibling, 1993; Molina et al., 1998; Greb and Archer, 2007; Le Heron and Etienne, 2005) to the migration of fluids during burial diagenesis (Davies et al., 2006). Among these processes, the seismic trigger remains the most commonly invoked for large-scale deformation and intrusion (Jolly and Lonergan, 2002; Owen and Moretti, 2011).

As our understanding of the controlling factors on the formation of intrusions and other soft-sediment deformation improves, it becomes possible to extract key information from their morphology, orientations, geographic distribution, and stratigraphic location (e.g. Obermeier, 1996; Boehm and Moore, 2002; Loope et al., 2013; Monnier et al., 2015). When coupled with detailed sedimentological and stratigraphic analyses, these features can considerably improve models for the structural setting during deposition and burial. A detailed understanding of sheet-like sedimentary intrusions, indicating sediment movement into planar fractures, may thus shed light on the distribution of pre- and syn-deformation fracture sets. Characterizing these fracture sets, by comparing them with known models of tectonic evolution and basement inheritance, aid in resolving the structural control on the preserved record.

In this study, previously undocumented sedimentary intrusions were investigated in the mixed siliciclastic-carbonate deposits of the Pennsylvanian Tensleep Formation (Northern Wyoming, USA). Based on their morphology, orientations, and distribution, a model for their formation is proposed. By comparing their orientation with documented attributes of the Archean/Proterozoic basement of North America, our interpretation suggests that the emplacement of these intrusions was controlled

predominantly by the configuration of the crystalline basement and the late Paleozoic geodynamic context in the western United States.

2. GEOLOGICAL SETTING

The Wyoming Shelf was an area characterized by low-accommodation rates during the late Paleozoic (Sloss, 1950). It was bounded to the west by the Antler Orogen and the associated foreland basin (a north-south oriented collisional mountain belt mostly active during the late Devonian and Mississippian; Poole, 1974; Dickinson et al., 1983). It was bounded to the north and south by the Big Snowy and Doughnut Troughs, respectively (Perry and Sloss, 1943; Willis, 1959; Rose, 1976; Maughan, 1984; Nelson and Lucas, 2011; Blanchard et al., 2015 and references therein). The origin of these higher accommodation domains is unresolved, but common explanations invoke the rejuvenation of basement structures (e.g. Shepard, 1993). The most significant tectonic event that affected the western United States during the late Paleozoic was the uplift of the Ancestral Rocky Mountains to the south-east of the Wyoming Shelf (Kluth and Conney, 1981; Kluth, 1986). The Ancestral Rocky Mountains are characterized by basement-cored, intracontinental blocks the uplift of which is thought to have been caused by the reactivation of pre-existing zones of weakness during the Ouachita-Marathon orogeny (Kluth and Conney, 1981; Kluth, 1986).

The Wyoming Shelf is located on the Wyoming Craton, an Archean continental fragment that accreted with its surroundings during the Paleoproterozoic (Whitmeyer and Karlstrom, 2007). This Archean craton consists of 2.9-2.5 Gyr-old granitic/granodioritic magmatic domains that alternate with 3.1-2.6 Gyr-old gneissic domain (Sims et al., 2001). This basement is thought to contain numerous lineaments, reactivations of which seemingly affected sedimentation throughout the Phanerozoic (Slack, 1981; Kraus, 1992; Inden and Coalson, 1996; Martinsen, 2003; Fielding et al., 2014). Two sets of lineaments are recognized in the Powder River Basin (Marrs and Raines, 1984): northwest-southeast

(N135-150°) and northeast-southwest (N55°). The continuity of these lineaments into the Bighorn Basin is, however, poorly constrained (e.g. Hoppin, 1974). The supercontinent Rodinia eventually broke up during the Neoproterozoic (Li et al., 2008). Some authors have attributed the existence of N to NE and W to NW trending fault-and-fold zones in the US Midcontinent to Mesoproterozoic stages of extension that predated the break-up of Rodinia (Marshak and Paulsen, 1996, 1997), leaving behind zones of weaknesses along basement rift faults.

This Archean craton was then covered by a thick succession of Paleozoic sediments, deposited during a time when the Wyoming Shelf was still an area of low-accommodation (Sloss, 1950). Pennsylvanian deposits mainly consist of alternations of sandstone and dolomite (Agatston, 1954). Each sandstone-dolomite couplet is laterally extensive (Carr-Crabaugh and Dunn, 1996; Blanchard et al., in revision), and can potentially be traced across the shelf. The Tensleep Formation, and its equivalents in neighboring states, were deposited as part of a large desert sand sea (or erg) that covered a substantial part of western Pangea during the middle Pennsylvanian (Steidtmann, 1974; Desmond et al., 1984; Driese and Dott, 1984; Kerr and Dott, 1988; Fig. 1A). This period was characterized by high-frequency fluctuations of sea-level, driven by orbitally-modulated waxing and waning of Gondwanan ice centers (Wanless and Shepard, 1936; Heckel, 1986; Davydov et al., 2010). This resulted in the deposition of cyclothemic alternations among limestone, coal and mudrocks with subordinate sandstones in the tropical realm (Heckel, 1986; Cecil, 1990), and the alternation of eolian dunes and marine dolomite in subtropical, arid regions (Steidtmann, 1974; Driese and Dott, 1984). In northern Wyoming, thick (3-15 m) eolian sandstones were deposited during sea-level fall (Carr-Crabaugh and Dunn, 1996) or lowstands (Blanchard et al., in revision; Fig. 2). They were subsequently preserved by an elevation of the water table that accompanied sea-level rise (Carr-Crabaugh and Dunn, 1996). Transgressive deposits consist of peritidal, mixed siliciclastic-carbonate facies overlain by shallow marine dolomite (Blanchard et al., in revision; Fig. 2). These facies are directly overlain by either peritidal or eolian sandstones, representing a significant basinward shift of facies and the development of a sequence boundary (Blanchard et al., in

revision; Fig. 2). These shifts in depositional environment are preserved as highly variable vertical stratigraphic trends in facies in which porous sands are overlain by laterally extensive, low-permeability marine dolomites.

Eolian sandstones, the hallmark of the Tensleep Formation, consist of cross-bedded, porous, very-fine to fine-grained quartz arenite (Agatston, 1952, 1954; Kerr and Dott, 1988). The combination of high porosity and fine grain size made these sandy units highly susceptible to liquefaction and fluidization (Obermeier, 1996), as evidenced by the presence of large-scale (several meters) convolutions of the original cross-bedding (Kerr et al., 1986). Some sandstone intervals were cemented soon after deposition by anhydrite and dolomite, due to the arid conditions and high evaporation rates that prevailed in this subtropical region (Mankiewicz and Steidtmann, 1979; Mou and Brenner, 1982). Early cementation of carbonate beds and peritidal sandstones suggest that they likely behaved in a brittle manner during burial.

3. METHODS

The Tensleep Formation is well exposed along the edges of the Bighorn Mountains in northern Wyoming, where it is commonly observable in cliffs and buttes. These topographic features provide access to outstanding 3D exposures. Sixteen of these localities were investigated (Fig. 1B). Out of these, four localities contained sedimentary intrusions and three had exposures that allowed measurement and sampling (White Creek Canyon, Dry Medicine Lodge Creek, Ziesman Canyon; Fig. 1B). Width, height, and orientation were systematically measured for a total of 103 intrusions. Where possible, diverse cliff orientations were surveyed to limit the bias introduced while collecting orientation data.

Internal structures within vertical sandstone bodies were recorded. Samples of intrusions were collected and thin sections made in order to characterize their petrography, relationships with the host rock, and potential sand sources, as well as the diagenesis that took place following formation.

The subsurface continuity of lithological units identified from studied exposures, and lateral changes in the thickness of units bounded by key surfaces, were analyzed using wireline logs from exploration drill holes in the Bighorn Basin, available from the Wyoming Oil and Gas Conservation Commission website (<http://wogcc.state.wy.us/>).

4. RESULTS

4.1 SPATIAL AND STRATIGRAPHIC DISTRIBUTION

Vertical, intrusive bodies were found in 3 locations (White Creek Canyon, Dry Medicine Lodge Creek, Ziesman Canyon; Fig. 1B). Similar, although thinner, sandstone dikes were also observed in a fourth locality (Webber Canyon; Fig. 1B), but their poor exposure precluded any significant measurements and descriptions.

Most intrusive bodies are found within 3 stratigraphic horizons (Fig. 3 and 4). In the northern exposures (White Creek Canyon), intrusive bodies are found in one horizon characterized by interbedded fossiliferous dolomite and purple shales. The abundance of shale interbedded with the dolomite beds of the intrusion-bearing horizon means that it generally tends to form recessive slopes. This lithological difference limits the exposure of quality rock faces through the horizon of interest. In the southern part of the study area (Dry Medicine Lodge Creek; Ziesman Canyon) intrusive bodies are found in two thick dolomitic intervals separated vertically by several meters (~3 m) of peritidal to eolian sandstones.

The localities where intrusions were observed are all located within a 23 km radius zone (Fig. 1B). Two other localities were visited within that zone (Alkali Flats, Trapper Creek), but the interval in which these features are found was not accessible and/or exposed. As such, the presence or absence of intrusive bodies at these localities could not be assessed.

Penetrative sedimentary intrusions are not the only type of soft-sediment deformation documented. Convolute laminations and bedding are pervasive in the Tensleep Formation. They are not limited to specific stratigraphic intervals, and are found at most exposures of the Tensleep Formation (Fig. 1). Small-scale (< 40 cm) convolutions of cross-bedding are in some cases truncated and overlain by undeformed sets of flat- or cross-stratified sandstones (Fig. 5).

4.2 MORPHOLOGY AND PETROGRAPHIC CHARACTERISTICS OF SEDIMENTARY INTRUSIONS

Penetrative, vertical sandstone bodies found in the Tensleep Formation are generally sheet-like (Fig. 6 and 7A, B, C). Their margins are discordant with the host rock (Fig. 7A, B, C), and their size is variable. Average height is 1 m, but ranges from 0.2 to 2 m. Width rarely exceeds 0.4 m, with an average of 0.25 m. Width can vary vertically within a single intrusion where it crosses different lithologies (Fig. 6 and 8). In such cases, intrusions are manifested as thin (1-5 cm) vertical streaks of sand within dolomitic beds, and they become thicker (up to 0.4 m) where the encasing lithology was less competent (Fig. 8). At Dry Medicine Lodge Creek (Fig. 1), intrusions found in the stratigraphically lower interval are smaller than those observed in the main unit (average height = 0.5 m; average width = 0.15 m). Due to the incomplete exposure of the length of intrusions, this dimension is not quantifiable from outcrops alone.

Intrusions are mostly composed of very fine to fine, moderately to well-sorted, quartz grains (Fig. 7D), but some also contain fragments of the host rock (Fig. 9F and G). Host rocks, either dolomite or shale, are locally cross-cut by thin (0.5 to 4 cm), mostly fibrous calcite veins (Fig. 9A, B). These veins are in many cases bedding-parallel, but some cross-cut the original bedding at a high-angle (> 45°). No cross-cutting relationships were observed between intrusions and calcite veins. These veins were only observed in association with sedimentary intrusions. Similar marine dolomite beds at other locations did not

192 contain any veins, nor did other marine intervals at locations where intrusions were studied. Host rock
193 beds are commonly deformed around the margin of intrusions. Beds are either upturned or downturned
194 (Fig. 6 and 7A, B, C). Many sandstone beds, or mixed siliciclastic-carbonate intervals are folded directly
195 below and above intrusions (Fig. 7A, B).

196 Based on geometry and degree of cementation, two main types of vertical intrusions can be
197 differentiated. The first type is characteristic of northern exposures (White Creek Canyon). These
198 intrusions are generally poorly cemented, and appear to taper downward (Fig. 9C, D, E). The top of these
199 intrusions is typically flat, and separated from the overlying eolian or peritidal sandstones by a layer of
200 purple shale (Fig. 9C, D). Thinner vertical bodies can be found extending downward from the thick part
201 of this type of intrusions (Fig. 9D). These thinner intrusions are typically ptymatically folded (Fig. 9D).
202 Fragments from the host shale are common in the main intrusive body (Fig. 9E, F, G). These fragments
203 have been plastically deformed, and contain thin streaks of quartz grains suggestive of microfractures
204 (Fig. 9F, G).

205 The second type of intrusion, found in more southerly parts of the study area (Dry Medicine
206 Lodge Creek; Ziesman Canyon), is very well cemented (Fig. 7D). Most of these well-cemented intrusions
207 clearly taper upward (Fig. 7A, B, C), but they also commonly appear to taper downward (Fig. 7A, B, C).
208 The top of these intrusions is often rounded. Flat tops have not been observed in this second type of
209 intrusions. Connection between surrounding sandstone bodies and intrusions is rarely observed. It is not
210 possible to assess with certainty whether this apparent contact reflects an original connection, implying a
211 genetic relationship between the intrusion and the horizontal body, or simply a secondary juxtaposition,
212 possibly reflecting complex geometries.

213 No evidence was found for ejection of sand at the water-sediment interface (i.e., sand volcanoes).
214 Sandstone intrusions are locally crossed by curved, slickensided surfaces that cut diagonally through
215 bodies (Fig. 10). These surfaces are limited to the intrusive bodies, and do not extend into the host rock.
216 Intrusion margins are offset where they intersect these slickensided surfaces.

217

218 4.3 ORIENTATION

219 The well-defined, sheet-like character of most of these intrusions allows the measurement of their
220 orientation. The preferential orientations of these intrusive bodies are consistent between the different
221 localities and beds where they were measured (Fig. 11). Two prominent directions are observed: N-S
222 (N005) and WNW-ESE (N115) (Fig. 11 and 12). A third, less pronounced set of intrusions oriented
223 WSW-ENE is also present (Fig. 11 and 12). This third set was apparently absent from the
224 stratigraphically-lower interval at Dry Medicine Lodge Creek (Fig. 11), although fewer data ($n = 15$) were
225 collected within this horizon. Some intrusion orientations are likely under-represented in locations where
226 the cliff face is oriented parallel to intrusions. The rose diagram representing the distribution of
227 orientations reflects this bias at White Creek Canyon, where the cliff is oriented parallel to the WNW-
228 ESE trending set (Fig. 11). As such, the WNW-ESE set does not appear prominent compared to the third
229 (WSW-ENE) set, which intersects the cliff face at a higher angle.

230 In some exceptional conditions, it is possible to observe the geometries and relationships between
231 these different sets where thin intrusions are exposed in the overhang of dolomitic beds (Fig. 8).
232 Intrusions oriented N-S abut against those oriented WNW-ESE (Fig. 8). The relationship between these
233 two sets and the third (WSW-ENE) has not been observed, due to its lesser abundance.

234

235 5. DISCUSSION

236 The sheet-like character of features suggests movement of material through open, planar fracture
237 planes that developed in semi-consolidated sediments. The source of these intrusions must have consisted
238 of non-cemented, very-fine grained quartz arenite, which are common in the eolian sandstones of the
239 Tensleep Formation. The upward-tapering of intrusions, and their continuation as thinner dikes extending
240 downward from the thicker part of intrusions are interpreted to reflect the upward movement of the

remobilized material. It is unclear whether the apparent downward-tapering character truly reflects downward movement or a significant lateral component to intrusion (Fig. 9, White Creek Canyon; c.f. Rodríguez-Pascua et al., 2000). The co-occurrence of downward- and upward-tapering in individual intrusions favors the latter interpretation (Fig. 6A, B, C). In the case of the northern variant (White Creek Canyon), apparent downward-tapering may also be explained by a thickening upward trend in upward-injected bodies. In the literature, sandstone dikes extending downward from source beds are more often than not interpreted as neptunian dikes (passive filling of open fractures by overlying sediments; Montenat et al., 2007), but it is worth mentioning that some authors have invoked downward injection of material to explain certain occurrences of downward-tapering intrusions (e.g. Parize et al., 2007; Ravier et al., 2015a). Downward injection of material remains a more unconventional mechanism, and is not considered a parsimonious explanation for the intrusions of the Tensleep Formation. A strong lateral component to sand injection (i.e. injection of fluidized sand from an underlying source bed in an upward direction, followed by a lateral flow through planar zones of weakness) is entirely consistent with observed morphologies.

Host rock fragments were excised during the process of intrusion. These fragments contain very thin ($< 200\ \mu\text{m}$) sand-filled microfractures, which are interpreted to result from micro-scale fracturing of margins during injection. Fragments of margins were then separated and transported within the fluidized sand mass (c.f. Ravier et al., 2015b).

The exclusive association between calcite veins and sandstone intrusions suggest that they were genetically related. Multiple occurrences of bedding-parallel calcite veins have recently been interpreted as resulting from the overcoming of vertical forces by fluid overpressure (Cobbold et al., 2013). If induced by hydrocarbon maturation (e.g. Cobbold et al., 2013; Zanella et al., 2014), the overpressure responsible for the generation of calcite veins should occur at great depths ($> 2\ \text{km}$). Considering the lack of any evidence for hydrocarbons within the veins, the presence of numerous veins cross-cutting bedding at high-angle, and their geographic distribution restricted to intrusion-bearing horizons, the previously

mentioned phenomenon (i.e. overpressure caused by hydrocarbon maturation) does not seem like a plausible explanation for their origin. One would instead expect these features to be more pervasive if their generation was linked to the burial history of these strata. It is thus possible that these features were instead opened during the local, fluidization-inducing overpressure event discussed in the following sections.

5.1 POST-EMPLACEMENT DEFORMATION

Both host rock and intrusions were affected by compaction during burial. Due to their competent nature, the mechanical response of intrusions to vertical stress differed from that of the enclosing dolomites and shales. Greater thickness reduction in the surrounding host rock resulted in folding of more competent layers above and below intrusions (Fig. 7A, B). The pillar-like character (i.e. vertically-oriented body, more resistant than its lateral surrounding) of vertical intrusions may also explain the apparent upward- and downward-drag of host rock beds around intrusion margin (Fig. 7A, B, C) (c.f. Rodríguez-Pascua et al., 2000; Parize et al., 2007). Compaction of wide intrusive sandstone bodies, which may have been cemented during shallow burial, was accommodated by the creation of internal, curved fault surfaces rich in slickensides (Fig. 10). Some intrusions contain several surfaces that have been cemented by silica. Cross-cutting relationships between cemented and slickensided surfaces suggest different generations of fault slip. Thinner intrusions developed ptigmatic folds in response to compaction.

5.2 SEISMICALLY-INDUCED INJECTION STRUCTURES?

Vertical sandstone dikes result from several processes:

288 - If porewater pressure surpasses the minimum compressive stress (σ_3) and tensile strength of the
289 confining layer, hydrofracturing occurs and fluidized sand is injected as sheet-like bodies oriented
290 perpendicular to the minimum compressive stress (Jolly and Lonergan, 2002). As such, their
291 orientation can be used to reconstruct the paleostress field during injection (e.g. Boehm and
292 Moore, 2002).

293 - Fluidized sand can be injected into pre-existing fractures if pore pressure is greater than the
294 normal stress (σ_n) exerted on the latter (Jolly and Lonergan, 2002). The orientation of such
295 features would theoretically reflect the paleostress field active before the time of injection.

296 - Fissures opened during extension can be filled by overlying material, resulting in the creation of
297 neptunian dikes (Montenat et al., 2007). They can be filled passively or by fluidized injection
298 (Montenat et al., 2007).

299 Based on their upward-tapering (Fig. 7A, B, C) and the presence of fragments derived from host
300 shale (Fig. 9E, F, G), the sedimentary intrusions observed in the Tensleep Formation were likely formed
301 by the injection of fluidized sand. Water-saturated sand liquefies when grain contact is lost, and pore-
302 water temporarily supports the sediments (Lowe, 1975). Fluidization can subsequently occur if the
303 supporting fluid flows at a rate high enough to counteract gravitational forces (Lowe, 1975). Two things
304 are required for this phenomenon to occur: (1) unconsolidated, porous sands overlain by a low-
305 permeability cap that allows the building of overpressure, and (2) a mechanism to shift the supporting role
306 from grains to the interstitial fluid. Dolomitic beds would have acted as barriers to vertical flow, allowing
307 the buildup of pore-water pressure.

308 Although numerous processes are known to increase pore-water pressure to a point sufficient to
309 induce soft-sediment deformation (Owen and Moretti, 2011), two phenomena are often cited as the main
310 cause responsible. The most commonly cited trigger is seismicity (Obermeier, 1996; Jolly and Lonergan,
311 2002). Cyclic shearing, induced by the propagation of seismic waves, results in a reorganization of

loosely packed sands (Lowe, 1975). During that stage, the sediment is no longer grain-supported and interstitial fluids become the supporting agent (Lowe, 1975). Rapid loading provides an alternative trigger, especially in settings prone to high sedimentation rate and gravitational failure (Moretti et al., 2001; Strachan, 2002; Oliveira et al., 2009).

The size and morphology of intrusions described in this study do not seem compatible with other phenomena known to produce soft-sediment deformation contemporaneous with deposition, such as storms (Martel and Gibling, 1993; Molina et al., 1998), waves (Dalrymple, 1979), and tides (Greb and Archer, 2007; Fan et al., 2014). Furthermore, the host strata must have been at least partially lithified to accommodate the opening of planar fractures.

While common in deep marine or deltaic environments, sudden sediment loading is not easily achieved around arid coastlines on gently dipping shelves. Furthermore, intrusions in the Tensleep Formation occur within a spectrum of facies, and most importantly, below different facies (peritidal and eolian sandstones). If overpressure was the result of rapid loading, consistent facies stacking (related to a sudden depositional event) would instead be expected. None of the enclosing facies can be attributed to high sedimentation rates or gravitational failure. As such, the hypothesis of rapid loading is not compatible with the slow processes at work during the deposition of the Tensleep Formation.

Another possible trigger is the addition of an overpressured fluid (Jolly and Lonergan, 2002). The hypothesis of overpressured fluids making their way to the center of the Wyoming Shelf is also unlikely, considering its structural configuration. Although documented in areas where high subsidence rates allow the potential generation of high fluid pressure and its migration up-dip or through faults, this phenomenon is not to be expected in northern Wyoming during the late Paleozoic. Contemporaneous, deeply subsiding basins are hundreds of kilometers away, and high subsidence rates in the area were only achieved during the latter half of the Mesozoic (Thomas, 1965; May et al., 2013). A slow build-up of pressure in sandstone beds caused by gradual burial may in some situations account for liquefaction and fluidization of sand. In the present situation, this argument is weakened by the non-random orientations and sheet-like

character of intrusions, suggesting injection in pre-existing weakness planes. Although there is doubt as to whether these pre-existing fractures were crossing the entire marine intervals or not, they would have potentially prevented a slow build-up of pressure by allowing fluid circulation across permeability barriers. They are, however, compatible with a seismogenic model, in which overpressure is rapidly generated and is not prevented by fluid circulation through fracture planes.

Considering their exclusive location within a definable zone, the existence of similar facies without intrusions within the lateral continuity of the same horizon or in stratigraphically-distinct intervals, and the fact that overlying facies differ from one location to another, the overpressure responsible for fluidization of sand in this case is best interpreted as the result of seismic activity. Large-scale convolutions and folds have also been observed in surrounding eolian sandstones (Kerr et al., 1986). The liquefaction of water-saturated sands required to explain this kind of soft-sediment deformation in eolian settings is also commonly interpreted to result from seismic shaking (Doe and Dott, 1980; Horowitz, 1982; Chan and Bruhn, 2014). However, deformed cross-bedding is much more widespread and as such may have been induced by a non-seismic trigger (Kerr, 1989). Some examples of convoluted cross-bedding show evidence for syn-sedimentary truncation (Fig. 5), pre-dating the deposition of overlying sediments. Considering the presence of such features in eolian sandstone stories formed before the deposition of intruded beds (Uppermost cross-bedded interval at Webber Canyon; Fig. 4), their non-zonal occurrence across the Bighorn Basin, and their presence in similar facies deposited contemporaneously across the Wyoming Shelf (Steidtmann, 1974; Doe and Dott, 1980; Desmond et al., 1984; Driese and Dott, 1984; Saperstone and Ethridge, 1984), it is suggested that these structures are not seismically-induced but autogenic.

The distribution of injectites in the Tensleep Formation provides a useful means of assessing the origin of this inferred seismicity. Based on studies of historical earthquakes and their impact on modern environments (Ambraseys, 1988; Obermeier, 1996), it is possible to estimate the magnitude of paleo-earthquakes using the distance from the hypothesized earthquake to the furthest location affected by

liquefaction. Liquefaction features found over a total distance of 45 km require a moment magnitude higher than 6 (Fig. 13). Although the generation of an earthquake of such magnitude seems at odds with the structural context of northern Wyoming during the late Paleozoic, the inferred presence of other structural features may explain this seismicity.

5.3 IMPACT OF BASEMENT LINEAMENTS

Located within the Wyoming Shelf, the hypothesized epicenter would have been far from any plate boundary (the closest one would likely have been 400 km east) and can only be explained in terms of intraplate seismicity. In such settings, rejuvenation of basement structures can produce the required seismicity (Johnston and Schweig, 1996; Jacobi, 2002).

The impact of such structures on the stratigraphic architecture of Paleozoic successions has to this day not been fully explored, although some workers have hypothesized a structural control on the orientation of incised valleys developed prior to the deposition of the Permian Phosphoria Rock Complex (Inden and Coalson, 1996).

Very thick (> 6 m) eolian sandstones, mostly deposited above Maximum Flooding Zone 10 (Blanchard et al., in review), are not preserved north of Webber Canyon but are consistently found south of Alkali Flats (Fig. 4). A similar shift is observed in the subsurface, where the top of the Tensleep Formation consists of thin (2-4 m) sandstone bodies (characterized by high density and neutron porosity values) alternating with dolomites (low density and neutron porosity values) in the northern part of the basin (Fig. 14B). The top of the Tensleep Formation is characterized by thicker (>8 m) sandstone bodies toward the south (Fig. 4 and 14B). Thickness changes in subsurface and surface exposures seem to align with the location of a tear fault on the eastern edge of the Big Horn Mountains (Fig. 14A). This tear fault may have been influenced by the presence of a pre-existing basement structure (Stone, 2003). Based on available observations, we suggest that these different manifestations in structures and stratigraphic

architecture are all located along a southwest-northeast oriented regional divide (Fig. 14A). Linear trends in the lateral variation in facies and bed thicknesses have been documented throughout the Mesozoic and Cenozoic successions of Northern Wyoming (Martinsen, 2003). These trends are attributed to spatial variations in accommodation caused by vertical movements along pre-existing basement faults (Martinsen, 2003). Numerous lineaments, inherited from Precambrian structures, are thought to have been re-activated several times, if not continuously, during the Phanerozoic (Slack, 1981; Kraus, 1992; Inden and Coalson, 1996; Martinsen, 2003; Fielding et al., 2014).

Basement lineaments are dominantly northeast-trending in the Powder River Basin (Slack, 1981; Marrs and Raines, 1984). A consensus is lacking for the Bighorn Basin, where their orientations are either considered west-east (Hodgson, 1965; Hoppin, 1974) or southwest-northeast (Marrs and Raines, 1984). The abrupt thickening of Tensleep sequences in the subsurface and the change in the nature of exposures line up with the Granite Ridge tear fault (northeastern Bighorn Mountains) (Fig. 14). The location and orientation of this trend is consistent with other structural features (Bighorn-Custer lineament of Marrs and Raines, 1984; Granite Ridge Tear Fault of Stone, 2003), and other faults and lineaments with similar orientations are still clearly noticeable from satellite imagery (Tongue River Lineament; Kearney Creek Lineament; Fig. 14).

It is proposed here that synsedimentary displacement along a southwest-northeast trending lineament was responsible for the spatial changes in subsidence that allowed the preservation of thick eolian sandstones to the south, or the orientation of incised valleys that removed such facies in the north. A link between basement lineament seismicity and the occurrence of soft-sediment deformation has previously been drawn in other studies, based on matching geographical distribution (Jewel and Ettensohn, 2004; Huuse et al., 2005).

5.4 IMPLICATIONS FOR THE DEVELOPMENT OF EARLY FRACTURE NETWORKS

The near-vertical sheet-like morphology of sandstone dikes reflect injection along planar fractures. These fractures may either have opened as a consequence of overpressure during injection (i.e. hydrofracturing; Jolly and Lonergan, 2002) or may have been formed before injection as the consequence of tectonic forces or basement inheritance. Hydrofractures open perpendicular to the minimum compressive stress, and as such are good indicators of the configuration of the stress field during the time of injection (Jolly and Lonergan, 2002; Boehm and Moore, 2002). This process is, however, not compatible with the distinct sets of orientations described in the present study (Fig. 11 and 12).

An alternative explanation (i.e. injection of sand into pre-existing fractures) is thus proposed to account for their varied orientations. These fractures sets (Fig. 12) can be interpreted either as reflecting two different stages of regional deformation (generating tectonic fractures paralleling the main compressive stress), or as the result of the reactivation of basement weaknesses that facilitates the generation of fractures. Orientations and abutting relationships between different sets of fractures are extremely well preserved in originally more competent beds. Where intrusions penetrated these more competent beds, they manifest as clearly defined, thin sandy vertical planes (Fig. 8). Abutting relationships between different sets of fractures, observed where the aforementioned thin sandy vertical intrusions are found (Fig. 8), suggest two discrete fracture-inducing events, resulting in two dominant orientations: N005 and N115 (Fig. 8). The older set (WNW-ESE) would have formed perpendicular to a minimum compressive stress oriented SSW-NNE. The stress field would have then shifted, resulting in a main compressive stress oriented North-South, causing the parallel opening of the second set of fractures (N-S). There were, however, no significant tectonic events occurring close to the study area during the late Paleozoic. The Ancestral Rocky Mountains were uplifted during the Pennsylvanian and Permian as a result of far-field forces from the Ouachita-Marathon orogeny (Kluth and Conney, 1981; Kluth, 1986). Both the direction of the associated main compressive stress and the distance from this orogenic event are incompatible with the generation of WNW-ESE or SSW-NNE-oriented fractures in northern Wyoming.

A third explanation for the generation of planar fractures is favored. There are similarities between the two dominant orientations of intrusions documented here and the orientation of Midcontinent fault-and-fold zones (Marshak and Paulsen, 1996, 1997). According to Marshak and Paulsen (1996, 1997), these zones were formed over Proterozoic rift faults, which were persisting zones of weakness throughout the Phanerozoic. It is proposed here that fractures in the dolomite horizons of the Tensleep Formation formed parallel to those intersecting zones trending N to NE and W to NW (Marshak and Paulsen, 1996, 1997). It is worth noting that one set of intrusion orientations could potentially have been created by hydrofracturing of low-permeability intervals while sand was contemporaneously injected through pre-existing fractures (c.f. Monnier et al., 2015). Although it is not necessary to explain all orientations of sandstone intrusions by a structural inheritance, the similarity between the pair of dominant orientations documented here and the Midcontinent fault-and-fold zones suggest that most dikes were formed by injection through pre-existing planes. A far-field effect of compressional stresses associated with the Ouachita-Marathon orogeny and concomitant uplift of the Ancestral Rocky Mountains would explain the reactivation of these basement structures.

5.5 TIMING OF INJECTION

Precisely constraining the timing of injection is difficult, but it is possible to give a rough estimate of their context within the paragenetic evolution of the Tensleep Formation. The downward- and upward- drag of host shales and dolomites, caused by the competent character of intrusions, suggests that these marine intervals were not yet compacted, or only slightly compacted, during injection. Significant compaction of carbonate mud typically occurs during the early stages of burial (up to 50% thickness reduction following burial under 200 m of overburden; Goldhammer, 1997). As such, it is difficult to conceive a model where intrusions formed at great depths (> 1 km) and host carbonates still achieved compaction rates of 30% after being intruded (Blanchard et al., in revision). This rate of compaction,

based on the folding of layers above and below intrusions, is a strong indicator that sandstone intrusions must have been in place before considerable burial (< 200 m).

The sheet-like character of these intrusions, suggesting injections through fracture planes, indicate that the host rock was cohesive. Carbonates commonly achieve cementation contemporaneously to deposition or during the early stages of burial (Shinn, 1969; Taylor and Illing, 1969). Considering these interpretations and the synsedimentary character of the fault, it is thus interpreted that injection occurred during the Pennsylvanian or Permian.

CONCLUSIONS

Previously undescribed vertical sedimentary intrusions from the Tensleep Formation (Pennsylvanian, Northern Wyoming) are investigated. The study of these carbonate-hosted, meter-scale sandstone bodies, and the model proposed for their emplacement, provide insights into the structural control exerted on the stratigraphic architecture of this mixed siliciclastic-carbonate succession:

(1) Observed over a 45 km transect, in three close stratigraphic horizons, these upward- and downward-tapering penetrative sedimentary intrusions are thought to have formed by a seismically-induced injection of fluidized sand in planar fractures (Fig. 15). A major lateral component to the injection is responsible for the joint occurrence of upward- and downward-tapering.

(2) The preferential orientations of these intrusions, and their abutting relationships when accessible, are interpreted as reflecting the organization of pre-injection fractures (consisting predominantly of WNW-ESE and N-S sets). Similarities between these two sets and the orientation of Midcontinent fault-and-fold zones associated with Proterozoic extensional basement faults (trending N to NE and W to NW) suggest that these fractures were likely generated parallel to these basement structures.

(3) These features are localized around a zone characterized by an abrupt lateral change in formation thickness. The orientation and location of this zone is consistent with other basement lineaments in the region, known to have affected facies belts and thicknesses throughout the Phanerozoic. It is thus proposed that seismic activity was caused by the reactivation of a basement lineament, which laterally affected accommodation and/or preservation of eolian sandstones.

Soft-sediment deformation structures are increasingly recognized as a key element in developing a thorough understanding of sedimentary systems. Similar features may have been overlooked in other records. Their presence in successions deposited in arid settings could be mistakenly identified as dessiccation crack infill where exposure is only partial. Thinner, smaller intrusions can sometimes be mistaken as rhizoliths in these settings. Key characteristics (tapering upward, non-random orientations, microfractures in host rock fragments) can help decipher the origin of true injection structures. When combined with detailed sedimentological and stratigraphic analyses, their recognition can shed light on structural controls during deposition and burial, and provide important insights on the timing and nature of reactivated basement structures.

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LIST OF FIGURES:

Fig. 1: (A) Stratigraphic framework and ages of Pennsylvanian deposits in the study area. Time scale and ages from Gradstein et al. (2012); (B) Location of exposures where carbonate-hosted, sandstone intrusions have been observed (Delineation of formations from Love and Christiansen, 1985). AF = Alkali Flats; BC = Bear Canyon; BI = Barry's Island; CC = Cottonwood Canyon; DA = Dayton; FS = Five Springs Creek; ML = Dry Medicine Lodge Creek; NB = Northeast of Bighorn Lake; SC = Shell Canyon; SM = Sheep Mountain; TC = Tensleep Canyon; TR = Trapper Creek Canyon; WB = Webber Canyon; WC = White Creek Canyon; WR = Wind River Canyon; ZC = Ziesman Canyon.

Fig. 2: Idealized depositional sequence resulting from one high-frequency fluctuation of sea-level. Peritidal and eolian sandstones are deposited during sea-level lowstand. The following sea-level rise enabled their preservation (by elevating the water table), and subsequently resulted in the deposition of peritidal sandstones and/or mixed siliciclastic-carbonate sediments. Marine dolomites were deposited during sea-level highstand. Their upper contact with the overlying lithologies (peritidal carbonates, mixed siliciclastic-carbonate, or sandstones) is generally sharp. Sandstone intrusions are found in marine dolomite and shale facies.

Fig. 3: (A) Measured section through the upper part of the Tensleep Formation at Ziesman Canyon (Fig. 1). Vertical arrow denotes stratigraphic position of the principal intrusion-bearing unit; (B) Exposure of the upper part of the Tensleep Formation. (C) Detail of a bed of marine dolomite containing numerous vertical intrusions.

Fig. 4: Interpretive transect through detailed measured sections through the Amsden and Tensleep Formations on the western side of the Bighorn Mountains (See Fig. 1B). Green stars denote locations of intrusions.

Fig. 5: Examples of convolute cross-bedding. Oversteepened to folded cross-beds are truncated by the overlying flat-stratified to cross-bedded set.

Fig. 6: Main characteristics of vertical intrusive bodies. Left side summarize characteristics of intrusions observed at White Creek Canyon (Fig. 1). Right side summarize characteristics of intrusions found in the southern exposures (Dry Medicine Lodge Creek and Ziesman Canyon, Fig. 1).

772

773 Fig. 7: (A, B) Vertical intrusive bodies in a dolomitic interval. Intrusions taper upward and downward.
774 Note folding of surrounding layers caused by differential compaction (white arrows); (C) Intrusive body
775 displaying the typical sheet-like geometry; (D) Whole thin section scan and (E) photomicrograph of a
776 well-cemented intrusion.

777

778 Fig. 8: 3D representation of the architecture of one intrusion and its interpreted relationship to pre-
779 existing fracture networks. Red squares represent plane of each photograph. Vertical variations in
780 intrusion reflect competency changes between layers. Overhangs allow the observation of abutting
781 relationships. Notice sidestepping geometries in some intrusions. Dikes oriented N-S abut against those
782 oriented WNW-ESE, interpreted as a reflection of discrete fracture-inducing deformation events.

783

784 Fig. 9: (A) Thin, bedding-parallel calcite vein in a shale interval within the main intrusion-bearing
785 interval; (B) Thicker veins adjacent to an intrusion in a dolomitic interval; (C) Apparent downward
786 tapering of an intrusion at White Creek Canyon; (D) Flat-topped intrusion with ptlygmatically folded,
787 thinner dikes at its base. Black and white stripes on Jacob's staff are each 10 cm long; (E) Plastically-
788 deformed fragments of host shale in intrusion. Black and white stripes on Jacob's staff are each 10 cm
789 long; (F) Whole thin section scan of a moderately to poorly cemented intrusion with numerous fragments
790 of host shale; (G) Some fragments contain quartz-filled microfractures (white arrow).

791

792 Fig. 10: (A) Intrusion containing curved, slickensided surfaces (white arrows) cutting diagonally through
793 its body. Hammer is 30 cm long; (B) Detail of one of these surfaces; (C) Slickensided surface in another
794 intrusion. Contact between the latter and its host rock can be observed on the left side (black arrow); (D)

Interpretive model for the generation of slickensided surfaces during compaction by means of normal faulting.

Fig. 11: Measured orientations of planar, vertical intrusions. Two dominant sets are found at every locality (WNW-ESE and N-S). The WNW-ESE set is likely under-represented at White Creek Canyon (WC), due to the cliff being oriented parallel to this orientation.

Fig. 12: Combined measurements of intrusions orientations gathered in this study. Two orientations dominate: WNW-ESE and N-S. A third set (WSW-ENE) is also present, but considerably less conspicuous.

Fig. 13: Relationship between earthquake magnitude and the furthest distance at which liquefaction occurs, based on empirical data from Ambraseys (1988) and Obermeier (1996). The domain on the right side of the curve correspond to zones beyond the maximum reach of liquefaction-generating seismic waves. Red dot denotes the position of the earthquake hypothesized in this study to explain the generation of intrusive features, and the corresponding magnitude required to explain their distribution.

Figure 14: (A) Inferred location of hypothesized basement lineament based on thickness and facies changes in the Tensleep Formation. (B) Subsurface, interpretive transect through the Amsden and Tensleep Formation in wireline logs. Note change in the preservation of the upper Tensleep Formation between wells MCP FEDERAL 1-19 and SAGEBRUSH 31-47.

817 Fig. 15: Interpretive model to explain the generation of vertical, sandstone intrusions in the Tensleep
818 Formation. (A, B, C) Pre-injection setting: (A) Simplified North-South cross-section showing
819 preservation of thick eolian sandstones at the top of the Tensleep Formation; (B) Interpreted distribution
820 of pre-injection fractures; (C) Development of fractures in competent beds; (D, E, F) Seismically-induced
821 injection of fluidized sand through pre-existing fractures: (D) Simplified North-South cross-section and
822 hypothesized location of epicenter; (E) Seismic activity along a basement lineament, thought to have
823 impacted the stratigraphy of the Amsden and Tensleep Fms., could have produced the seismic waves
824 required to explain the fluidization of water-saturated eolian sands; (F) Seismic shaking causes a loss of
825 grain contact and an increase of porewater pressure. The resulting fluidized sand flows and gets injected
826 in pre-existing fractures.

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