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# Do soft sediment deformations in the Late Triassic and Early Jurassic of the UK record seismic activity during the break-up of Pangea?

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Marine Laborde-Casadaban, Catherine Homberg, Johann Schnyder, Sandra Borderie, Robert Raine. Do soft sediment deformations in the Late Triassic and Early Jurassic of the UK record seismic activity during the break-up of Pangea?. *Proceedings of the Geologists' Association*, 2021, 132 (6), pp.688-701. 10.1016/j.pgeola.2021.02.007 . hal-03625737

**HAL Id: hal-03625737**

**<https://hal.science/hal-03625737v1>**

Submitted on 8 Jan 2024

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37 2005; Greb and Archer, 2007; Ghosh et al., 2012); lacustrine (Marco et al., 1996; Ken-Tor et al., 2001)  
38 and deep marine – turbidite environments (Allen, 1977; Bergerat et al., 2011; Homberg et al., 2013;  
39 Basilone et al., 2015). These deformations include convolute bedding, overturned cross-stratification,  
40 load structures or water escape features. Due to their low cohesion and small grain size with large  
41 water content, silts are particularly sensitive to liquefaction (Allen 1977). During liquefaction, the  
42 pore-fluid pressure increases and the grain weights are transferred to the fluid. This causes the loss  
43 of grain packing and thus of the internal sediment organisation (Lowe, 1975, 1976; Allen, 1977). The  
44 association of liquefaction with a driving force, such as gravity, an unstable density gradient or a  
45 shear stress leads to the deformation of the original sedimentary deposit (Owen, 1987). Numerous  
46 triggers may cause the liquefaction of sediments such as: (1) tsunamis (Takashimizu and Masuda,  
47 2000; Nanayama et al., 2000; Schnyder et al., 2005; Le Roux et al., 2008), (2) tidal flux and tidal bores  
48 (Tessier and Terwindt, 1994; Greb and Archer, 2007), (3) storm waves and breaking waves (Molina et  
49 al. , 1998) or (4) rapid sedimentation and loading (Allen, 1982). **Earthquake(s) as a trigger of the**  
50 **observed SSDs is frequently proposed in the literature**, yet in many cases the liquefaction is  
51 attributed to a seismic shock without sufficient evidence (Pope et al., 1997; Kullberg et al., 2001).  
52 This shows the difficulty in identifying earthquakes as the cause of sediment liquefaction without  
53 well-defined and self-sufficient criteria. Several authors have shown that assigning earthquakes as  
54 causing SSDs requires a number of criteria (Wheeler, 2002; Owen and Moretti, 2011), including:  
55  
56 (1) The occurrence of liquefied sediment layers,  
57 (2) A large extent of synchronous liquefied event,  
58 (3) A lateral continuity of a deformed bed along the outcrop,  
59 (4) Similar structures to those observed during recent earthquakes (Audemard and De Santis, 1991;  
60 Dechen and Aiping, 2012),  
61 (5) The superposition of SSD levels in the sedimentary pile which may correspond to a succession of  
62 seismic events. This criterion is especially strong when it is possible to link the deformed levels to  
63 historical earthquakes (Marco et al., 1996; Ken-Tor et al., 2001),  
64 (6) A sedimentary and tectonic context consistent with the occurrence of frequent earthquakes  
65 (Bergerat et al., 2011).

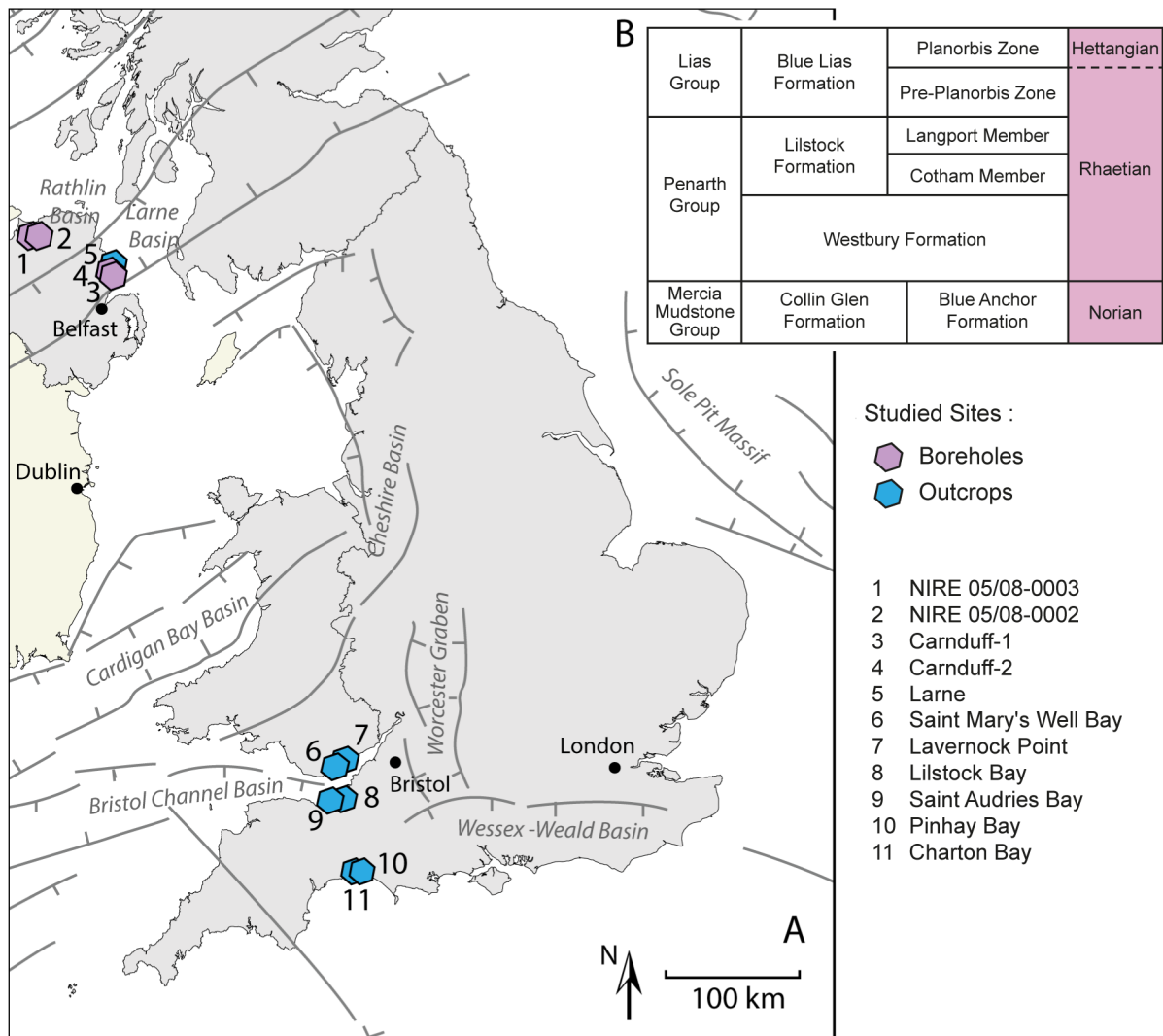
66 Nevertheless, none of these criteria are direct evidence of earthquakes as a trigger. Indeed, very few  
67 studies show the direct relationship between syn-sedimentary faults and SSDs (Basilone et al., 2015).  
68 The Triassic-Jurassic boundary strata from the UK are affected by numerous soft deformation  
69 structures recorded at a large scale (Richardson, 1911; Hallam, 1960; Mayall, 1983; Gallois, 2005,  
70 2007, 2009). Several authors (Mayall, 1983; Hesselbo et al., 2002, 2004; Simms, 2003, 2007) assigned

71 those SSDs to earthquakes but the number of events and their origin are still debated. This study  
72 investigates how the orientation of various SSDs are likely to inform on the SSDs trigger in Triassic–  
73 Jurassic boundary strata from Southern England, South Wales and Northern Ireland. In the following,  
74 we combined sedimentological and structural approaches to investigate the geometry, cross-cutting  
75 relationships and distribution of the observed SSDs (faults, folds, slumps, domes), in order to assess  
76 the possible earthquake trigger of the disturbed layers and to discuss the relationships of these  
77 deformations with the Triassic–Jurassic rifting context.

78

## 79 **2. Geological context at the Triassic-Jurassic boundary in the UK**

80 The break-up of the supercontinent Pangea resulted in the opening of the Neotethys Ocean to the  
81 east during the Late Permian – Early Triassic times (Ziegler and Stampfli, 2001). In north-western  
82 Europe, numerous sedimentary basins related to this large-scale extensional tectonic context, and  
83 often bounded by reactivated Palaeozoic faults, formed during the Triassic-Early Jurassic (Ziegler and  
84 Dezes, 2006). This is the case for the N–S Worcester Basin, the E–W oriented Bristol Channel and  
85 Wessex basins and the ENE-WSW oriented Lough Foyle and Larne basins (Fig. 1A, Chadwick, 1993;  
86 Holdsworth et al., 2012). Subsidence of these intracratonic basins continued during Jurassic and  
87 Cretaceous times.



88  
89 Figure 1. Tectonic framework of the UK at the Triassic–Jurassic transition. A: Location of studied sites.

90 The faults which mark the different Permo-Triassic basins are those defined by Holdsworth et al.  
91 (2012). B: Lithostratigraphy of the Triassic-Jurassic boundary deposits in the UK. After Warrington et  
92 al. (1994) and Mitchell (2004).

93 The sediments preserved in the UK basins recorded a long-term transgressive period extending from  
94 the Late Triassic until the Early Jurassic (Warrington et al., 1980), starting with the continental  
95 floodplain and lake deposits of the Mercia Mudstone Group of Triassic age. In England, the upper  
96 unit of the Mercia Mudstone Group corresponds to the green to grey mudstones of the Blue Anchor  
97 Formation, and in Northern Ireland to the red and green silty mudstone of the Collin Glen Formation  
98 (Mitchell, 2004). Ubiquitously overlying the Mercia Mudstone Group is the relatively thin Penarth  
99 Group. Most of the SSDs from SW England, South Wales and Northern Ireland are confined to the  
100 Penarth Group (Rhaetian, Late Triassic), which is the primary focus of this study. The Penarth Group  
101 can in most instances be subdivided into the Westbury Formation and the overlying Lilstock  
102 Formation (Cotham Member and overlying Langport Member, Fig. 1B).

103 The Westbury Formation comprises dark-grey, silty, laminated mudstone and sandstones with  
 104 current and wave ripples. It has yielded bioclastic accumulations of bivalve fragments, fish bones of  
 105 marine and semi-aquatic origin and locally few terrestrial vertebrate remains (Ivimey-Cook, 1974;  
 106 Storrs, 1994; Radley and Carpenter, 1998; Swift and Martill, 1999). It is interpreted to be deposited in  
 107 lagoonal to shallow marine environments (Hesselbo et al., 2004). In the Severn Estuary area (Bristol  
 108 Channel Basin), the top of the Westbury Formation is more clay rich. It is intercalated with nodular  
 109 muddy or shelly limestone and sandstones (Richardson, 1911; Radley and Carpenter, 1998). The  
 110 overlying Lilstock Formation is divided into two constituent members. At its base, the Cotham  
 111 Member corresponds to silty-sandy laminated grey-green mudstones with wave ripples. It has been  
 112 interpreted as a coastal, shallow marine to freshwater lagoon environment (Mayall, 1983; Radley and  
 113 Swift, 2002; Gallois, 2009). In the Bristol Channel area, a distinctive level of so-called desiccation  
 114 cracks with *per-descensum* clastic dykes separates the lower Cotham Member from the upper  
 115 Cotham Member (Ivimey-Cook, 1974; Mayall, 1983; Hesselbo et al., 2002; Gallois, 2009). A similar  
 116 horizon was recognised by Simms (2003, 2007 and Simms and Jeram, 2007) from the Waterloo  
 117 section in Northern Ireland. Although it may correlate with the horizon in SW Great Britain, it sits at a  
 118 higher stratigraphic level in the Cotham Member (Jeram et al., this volume). **These cracks have been**  
 119 **recently re-interpreted as subaqueous sedimentary (so-called syneresis) cracks (Jeram et al., this**  
 120 **volume).** The upper part of the Cotham Member marks a clear enrichment in sand at all localities. At  
 121 the top of the Penarth Group, the Langport Member is usually a blue grey to very light grey  
 122 mudstone and limestone unit. In SW England, where the very light grey micritic limestones are more  
 123 common, they have alternatively been referred to as the White Lias. In SW England, the Langport  
 124 Member is also associated with numerous gravitational transport process (Hallam, 1960; Wignall,  
 125 2001; Hesselbo et al., 2004). It is particularly well-developed on the South Devon coast where it is  
 126 around eight metres thick. The member is restricted to a few decimetre-thick beds in the Severn  
 127 Estuary area but in Northern Ireland it is between 4.92 m and 9.27 m thick (Raine et al., this volume).  
 128 These deposits correspond to lagoonal to fully marine conditions (Richardson, 1911; Swift, 1995).  
 129 Across much of southern Great Britain, the Langport Member is capped by a layer penetrated by  
 130 *Diplocraterion* burrows indicating erosion and sediment-starvation in a bed referred to as the “Sun  
 131 Bed” (Wignall, 2001; Radley and Swift, 2002; Hesselbo et al., 2004). Finally, the succeeding dark-grey  
 132 mudstone–limestone alternations of the Blue Lias Formation in Great Britain and the mudstone  
 133 dominated Waterloo Mudstone Formation in Northern Ireland (Hettangian, Early Jurassic) attest to a  
 134 deeper marine environment (Deconinck et al., 2003; Gallois, 2007). The stratigraphical nomenclature  
 135 used in our logged sections are those used by Mitchell (2004) for Northern Ireland, and by  
 136 Warrington et al. (1980) for SW Great Britain.

### 137 3. Methods and data

138 Seven outcrops and four boreholes spanning Late Triassic – Early Jurassic strata were studied across  
139 Northern Ireland and Great Britain (Fig. 1A). They encompass several Permo-Triassic basins: the  
140 Larne and Lough Foyle basins in Northern Ireland and the Wessex and Bristol Channel basins in the  
141 south and south west of Great Britain. Outcrop observations were supplemented by analysis of four  
142 cores drilled in Northern Ireland in the Larne Basin (Carnduff-1 and Carnduff-2) and in the Lough  
143 Foyle Basin (NIRE 05/08-0002 and NIRE 05/08-0003) (Fig. 1A).

144 The eleven selected cores and outcrop sections were logged in detail to identify the lithological units  
145 and the sedimentary features. In all cores and sections, we observed deformation of the beds at  
146 various scales (SSDs), from convolute bedding, mesoscale folds, slumps, micro- and meso-scale faults,  
147 and small and large bodies of injected liquefied sediments. In some cases, the upward movement of  
148 the liquefied sediments resulted in elongated domes drawn by the upper limit of the liquefied bed.  
149 Data collected on the SSDs include (1) their vertical distribution within the sedimentary pile, (2)  
150 orientation (strike and dip) and maximum offset of the faults; (3) direction of fold axis and of their  
151 overturning direction and (4) strike of elongated domes. The good quality of the outcrops allowed to  
152 collect numerous measurements, so that the direction of the SSDs is accurately constrained in each  
153 studied site. Attention was paid to identify lateral variations in the SSD trends in each level and  
154 variations from one bed to another within a section. Orientation data include 360 measurements in  
155 Pinhay Bay, 31 measurements in Lavernock, 34 measurements in Lilstock, 31 measurements in Saint  
156 Audrie's Bay, 91 measurements in Waterloo Bay. No direction data could be obtained in the  
157 Northern Ireland cores because they were not oriented. These data together allow to characterize  
158 the recurrence and style of deformation in relation with the lithology and to obtain a statistically  
159 representative determination of the SSD orientations.

160

## 161 4. Results: distribution and directional characterisation of SSD in Triassic–Jurassic boundary

### 162 *strata*4.1 Northern Ireland (Larne and Lough Foyle basins)

#### 163 4.1.1 Distribution of SSDs

164 On the Waterloo Bay foreshore section at Larne, the Westbury Formation corresponds to dark grey  
165 laminated clays with occasional silty to sandy laminated layers (Figs. 2 and 3). The boundary between  
166 the Westbury Formation and the Cotham Member is deformed by centimetre- to decimetre-scale  
167 dome structures (Figs. 2 and 3B). This is particularly visible in the western part of the outcrop. The  
168 SSDs affect units 1 to 6H (Fig. 2). Dark grey mudstone of the upper Westbury Formation is injected

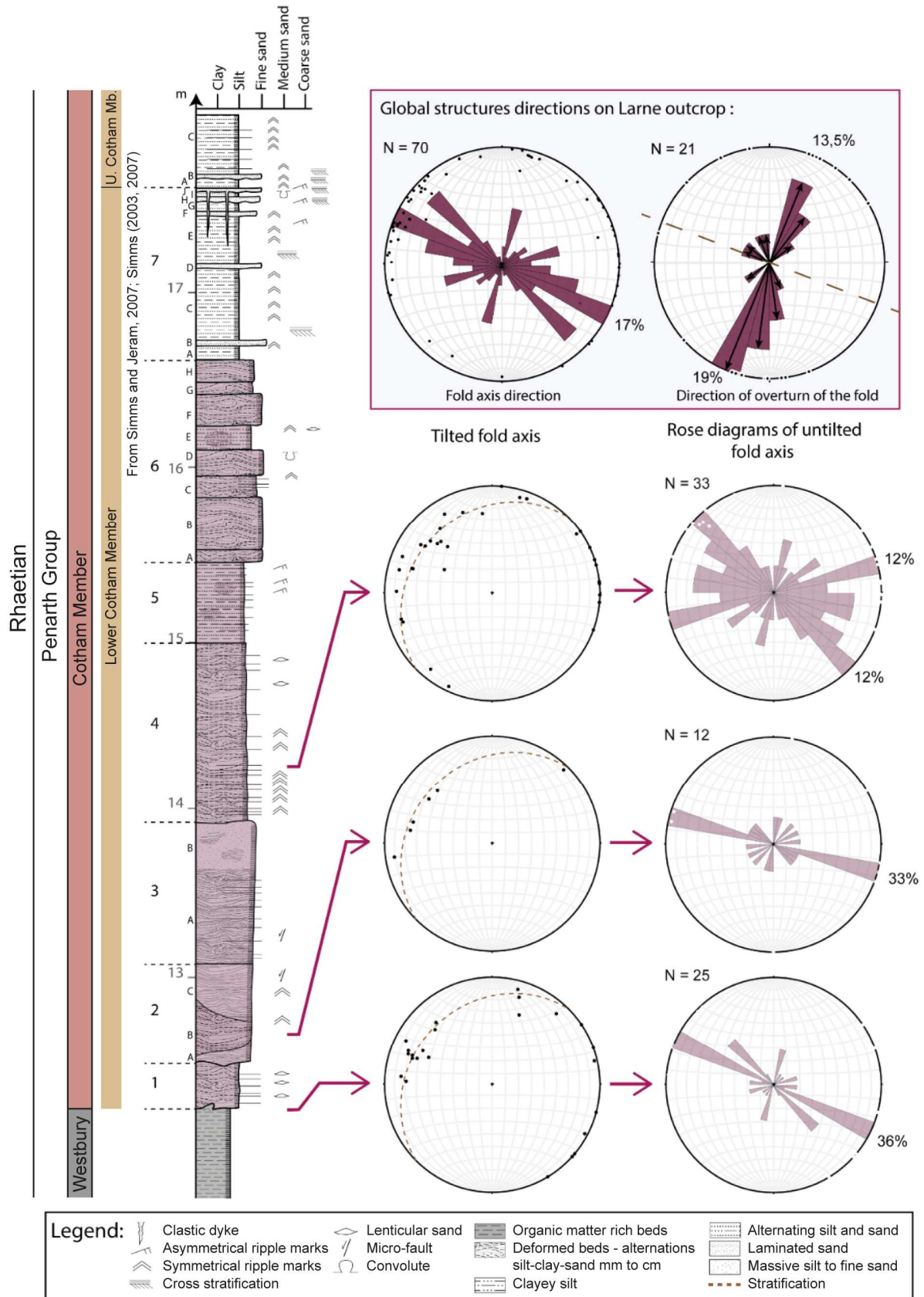
169 throughout the overlying coarser sediments of the Cotham Member, which pinch out against the  
170 dome walls. Claystone and sandstone alternations of the basal Cotham Member are deformed  
171 around these injections and form folds caused by the upward movement of the clay. These  
172 deformations include diapirs and anticlinal cusps corresponding to fluid escape from overpressured  
173 clay up through the non-indurated sandy alternations (Ghosh et al., 2012). The lower Cotham  
174 Member is furthermore disturbed by numerous centimetre- to metre-scale isoclinal and recumbent  
175 folds (Fig. 3C and D). These folds affect laminated millimetre- to centimetre-thick beds of siltstone,  
176 sandstone and claystone (units 1 to 6, Fig. 2). The larger folds are in the lenticular unit 2B and have a  
177 width and height of 1.0–1.5 m. Second-order folds, with shorter wavelength are observed in the  
178 sandier intervals (Fig. 3D). Some areas are stretched and others contracted causing boudinage and  
179 variations of sandstone layer thickness, which has amplified the shape of existing wave ripples (Fig.  
180 3D). The interval with SSDs is 4.6 m in thickness. However, the density of SSDs varies laterally, and  
181 less deformed areas can be recognised. Unit 3B corresponds to a finer and more homogeneous  
182 siltstone that was less affected by soft deformations. From unit 7 to the top of the section, the layers  
183 are undisturbed despite having an overall similar lithology to the underlying deformed units (Fig. 2).  
184 Only one level with *per-descensum* centimetre-thick dykes was recognised at the top of unit 7. **These**  
185 **dykes show multidirectional orientations and were first recorded by Simms and Jeram (2007). They**  
186 **are discussed in detail and interpreted by Jeram et al. (this volume) as two or three generations of**  
187 **subaqueous sedimentary (so-called syneresis) cracking, possibly linked to a seismic event.**

188 The Carnduff-1 and Carnduff-2 cores from the Larne Basin are 2.82 and 2.58 km away, respectively,  
189 from the Waterloo Bay section at Larne (Fig. 4). The cores display very similar features to those of the  
190 outcrop. Thicknesses of the individual lithostratigraphical units are comparable between the cores  
191 and the nearby outcrop. SSDs occur in the cores from the base of Cotham Member and comprise a  
192 4.3 m thick interval in Carnduff-1 and a 4.4 m thick interval in Carnduff-2 (Figure 4). The layers  
193 bearing the SSDs in these cores and in the Waterloo Bay outcrop are easily correlated using marker-  
194 beds thanks to their limited geographical separation. The shell-rich levels affected by convolute  
195 bedding in the upper Westbury Formation, the three sandy layers at the top of Cotham Member, and  
196 the levels rich in bivalve shells at the base of Blue Lias Formation are very good local bed markers  
197 (Fig. 4). The deformations observed in Carnduff-1 and Carnduff-2 cores include millimetre- to  
198 centimetre-scale recumbent folds, isoclinal folds and convolute bedding. Numerous normal micro-  
199 faults with length of several centimetres also affect the upper half of the deformed level. This is well  
200 seen in finely laminated deposits that form micro-graben-like structures. Variations in bed thickness  
201 are observed on either side of several micro-faults, suggesting that some of them are syndimentary



202 faults. The deposits above the deformed level in the Carnduff-1 core are cut by a centimetre-wide  
203 clastic dyke.

204



205

206 Figure 2. Detailed log section of the Waterloo Bay foreshore, Larne. SSDs are observed in units 1 to 6  
 207 (beds in purple). Plots on the left correspond to fold axes with the present day dip. Rose diagrams on

208 the right correspond to the direction distribution of folds along the outcrops. Both diagrams at the  
209 top give the fold axis distribution for the whole section. N: number of data. Note that all fold axes lie  
210 in the bedding plane. Folds thus formed with a sub-horizontal axis before the tilting of beds.

211 The NIRE 05/08-0002 and NIRE 05/08-0003 cores are situated in the Lough Foyle Basin in Co.  
212 Londonderry, about 68 km NW from the Waterloo Bay section (Figs. 1 and 4). In the Lough Foyle  
213 Basin, the tidal (flaser-bedded) deposits of the Cotham Member are thicker than those of the Larne  
214 Basin. The base of the formation is characterised by a massive, apparently un-deformed grey-green  
215 silty clay. Elsewhere in the region this grey clay has been recorded as having *Rhaetavicula contorta*, a  
216 common bivalve in the Westbury Formation (Bazley et al., 1997). The Cotham Member in the Lough  
217 Foyle Basin comprises comparable facies and numerous recumbent and isoclinal folds with similar  
218 size to those in the Cotham Member in the Larne Basin, as observed in the Waterloo Bay section and  
219 in Carnduff-1 and Carnduff-2 cores. The deformed level in NIRE 05/08-0003 is 5.8 meters in thickness  
220 (Fig. 4). Numerous synsedimentary normal micro-faults are observed throughout the sandy layers of  
221 the deformed level. In NIRE 05/08-0002, deformations are restricted to a two-metre-thick interval in  
222 the upper part of the Cotham Member. However, because of the condition of the core, it cannot be  
223 determined whether there are deformations in the lower part. In Northern Ireland, the original  
224 laminated internal structure of the deposit is always preserved during the deformation. The  
225 widespread occurrence of overturned folds suggests that a shear stress has been applied on the  
226 sedimentary material during liquefaction of the sediments.

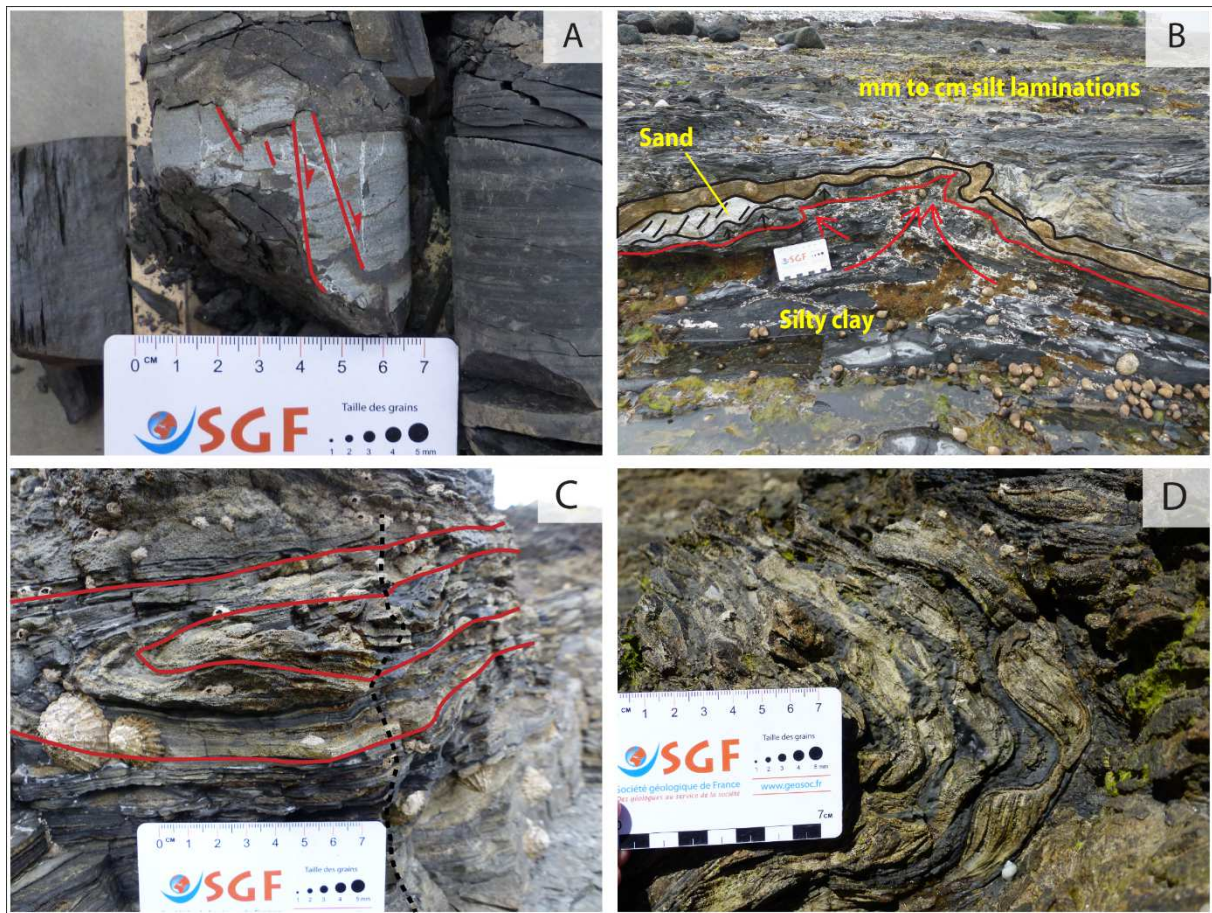
227 To summarise, the same disturbed interval of the lower Cotham Member in Northern Ireland (level 2,  
228 Fig. 4) can be correlated between the 5 sections throughout the two basins.

229 In addition to this main deformed layer in the Cotham Member, the base of the Westbury Formation  
230 is also affected by soft deformations in the NIRE 05/08-0002 and NIRE 05/08-0003 cores (Fig. 4). The  
231 thickness of these deformed levels (level 1, Fig. 4) is respectively 2.5 and 1.0 m. Within those two  
232 cores, the deformations are especially well-observed in a 15 to 20 cm thick interval made of clay and  
233 sand laminae. Some of the centimetre-scale folds, convolute bedding, sandstone boudins and  
234 angular undeformed clasts may result from bioturbation. However, the occurrence of several  
235 millimetric to centimetric normal faults and graben structures throughout the two cores in the  
236 Westbury Formation indicates that the deformations are most likely non-biological in origin (Fig. 3A).  
237 Thickness variations from either side of the observed micro-faults also indicate syn-sedimentary  
238 activity during the deposition of the Westbury Formation.

239 Sand- and shell-rich levels disturbed by convolute bedding and synsedimentary faults of millimetre-  
240 to centimetre displacement in the Westbury Formation are interpreted as a moderate deformation

241 event (level 1, Fig. 4) that affected the Westbury Formation in the Lough Foyle Basin. In our studied  
242 sites in the Larne Basin, some convolute bedded shelly levels with burrows are found at the top of  
243 the Westbury Formation (Fig. 4), and are likely to equate to this deformation event, as other SSDs  
244 have been found in the upper half of the Westbury Fm. in Cloghfin Port, South of Islandmagee  
245 (Jeram, et al. this volume). However, the absence of normal faults and of typical liquefaction  
246 structures does not rule out the possibility of a bioturbation origin.

247



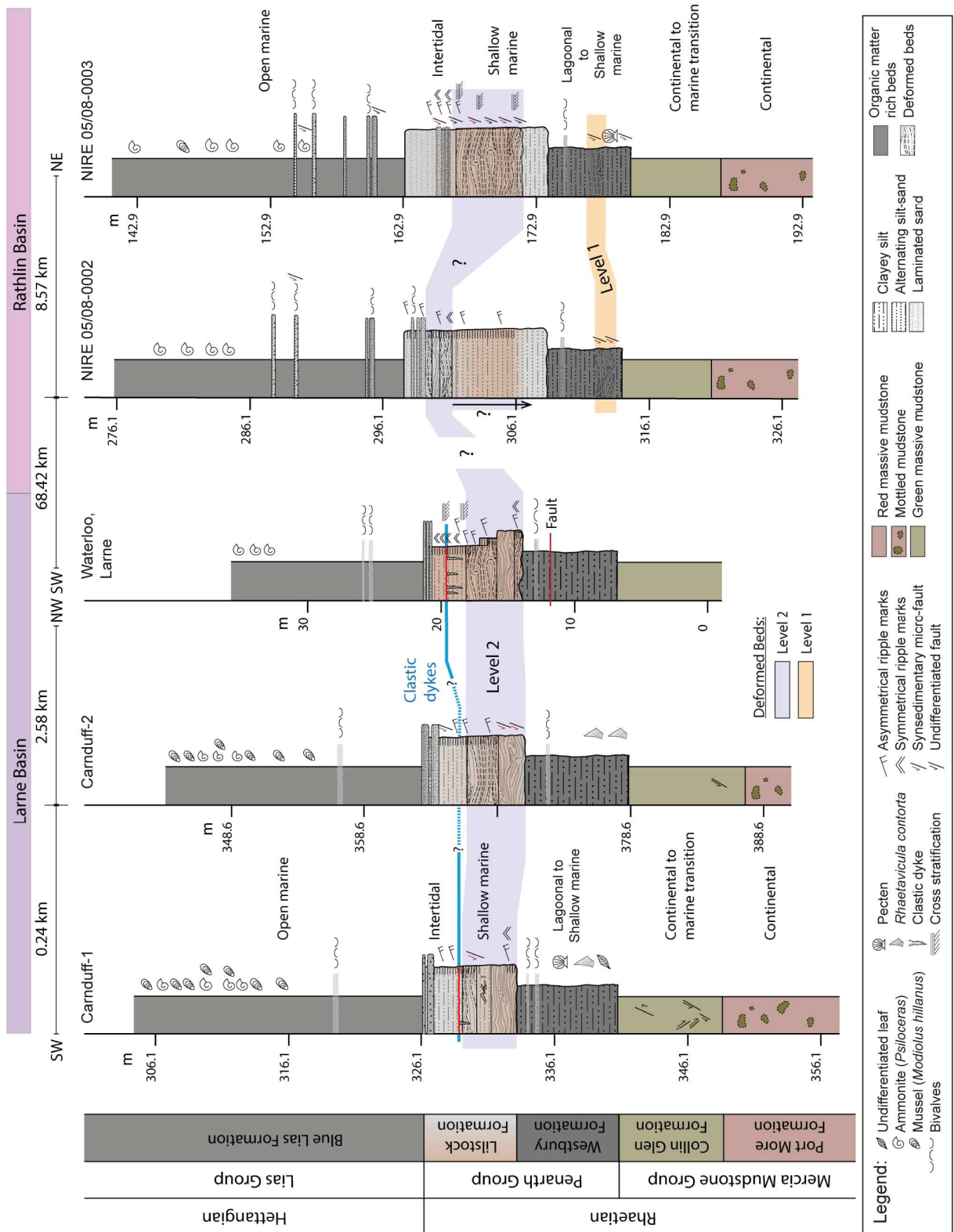
248  
249 Figure 3. SSDs observed in Northern Ireland. (A) Synsedimentary micro-faults in the Westbury  
250 Formation, NIRE 05/08-0003 core. (B) Diapir injection structure at the boundary between the  
251 Westbury Formation and the Cotham Member, Waterloo Bay. (C) Recumbent fold in the lower  
252 Cotham Member, Waterloo Bay. (D) Second generation of folds and boudinage in a larger fold,  
253 Waterloo Bay.

254

255

256





257

258 Figure 4: Log sections and position of deformed layers in Northern Ireland. Carnduff-1, Carnduff-2,  
 259 NIRE 05/08-0002 and NIRE 05/08-0003 correspond to cores, and the Waterloo, Larne, corresponds to  
 260 the outcrop in Waterloo Bay foreshore.

261 4.1.2 Directional characterisation of the SSDs (Waterloo Bay foreshore, Larne Basin)

262 The excellent outcrop at the Waterloo Bay foreshore exposure in Northern Ireland, allows the SSD  
263 direction of the SSDs to be studied. At the boundary between the Westbury Formation and the  
264 overlying Cotham Member, the elongation of the diapirs show a clear preferred N120°E orientation,  
265 accounting for 36% of the 25 measured injections (Fig. 2). In unit 2B, various fold orientations have  
266 been observed but a large proportion (33% of 12 measurements) show a roughly similar (N110°E)  
267 trend. Fold directions are more scattered in unit 4 where two main directions are recognised: 12%  
268 have a N130°E direction, similar to that observed in underlying layers and 12% have a N70°E  
269 direction from 33 measurements. These results highlight a mean NW–SE fold axis direction through  
270 the deformed levels. The overturned directions of the folds are either oriented toward the North or  
271 toward the South. This likely excludes sliding along local submarine slopes as a control of the  
272 observed soft-sediment folding. On the other hand, the constant orientation of the strain ellipse  
273 suggests a tectonic origin of the deformation structures that were thus controlled by the local state  
274 of stresses.

## 275 *4.2 Bristol Channel Basin*

### 276 *4.2.1 Distribution of SSDs*

277

278 In the Bristol Channel Basin, the Cotham Member corresponds to a calcareous mudstone with  
279 millimetre- to centimetre-scale siltstone and fine sandstone laminations. The lithology is substantially  
280 similar to Northern Ireland, with a more prevalent carbonate component. Six stratigraphic intervals  
281 have been defined at St Mary's Well Bay, eight at Lavernock Point, five at Lilstock, and five at St  
282 Audrie's Bay (Fig. 5). Lithofacies correlations using marker beds such as dyked horizons appear  
283 obvious between the four sections (Fig. 5). The boundary between the Westbury Formation and the  
284 Cotham Member is not affected by diapiric injections as in the Larne Basin. However, at Lavernock  
285 Point, recumbent folds involving pockets of dark organic-rich material with bioclasts that have  
286 originated from the underlying Westbury Formation are present in the first 70 cm of the Cotham  
287 Member (Fig. 5, bed 2A, and Fig. 6A). Reworked bivalves were recorded in the region in the basal  
288 beds of the Cotham Member by Waters and Lawrence (1987).

289 At St Mary's Well Bay, South Wales, and St Audrie's Bay, Somerset, the lower Cotham Member is also  
290 affected by numerous recumbent and isoclinal folds (Fig. 5 and Fig. 6C). At Lilstock, two distinct levels  
291 of deformation (N1 and N2, Fig.5) within the lower Cotham Member can be followed along the cliff  
292 and on the foreshore. The Lilstock section shows many overturned and recumbent folds in unit 2C (in  
293 the middle of the member) and in unit 2E (Fig. 5 and Fig. 6B). The two liquefied levels are here  
294 separated by an un-deformed 60 cm thick mudstone with beds of siltstone and fine sandstone (unit  
295 2D).

296 At St Audrie's Bay, St Mary's Well Bay and Lavernock Point, deformations are systematically located  
297 below the level of polygonal dykes (Fig. 5). At Lavernock point, the thickness of the disturbed interval  
298 is greater than at any other localities.

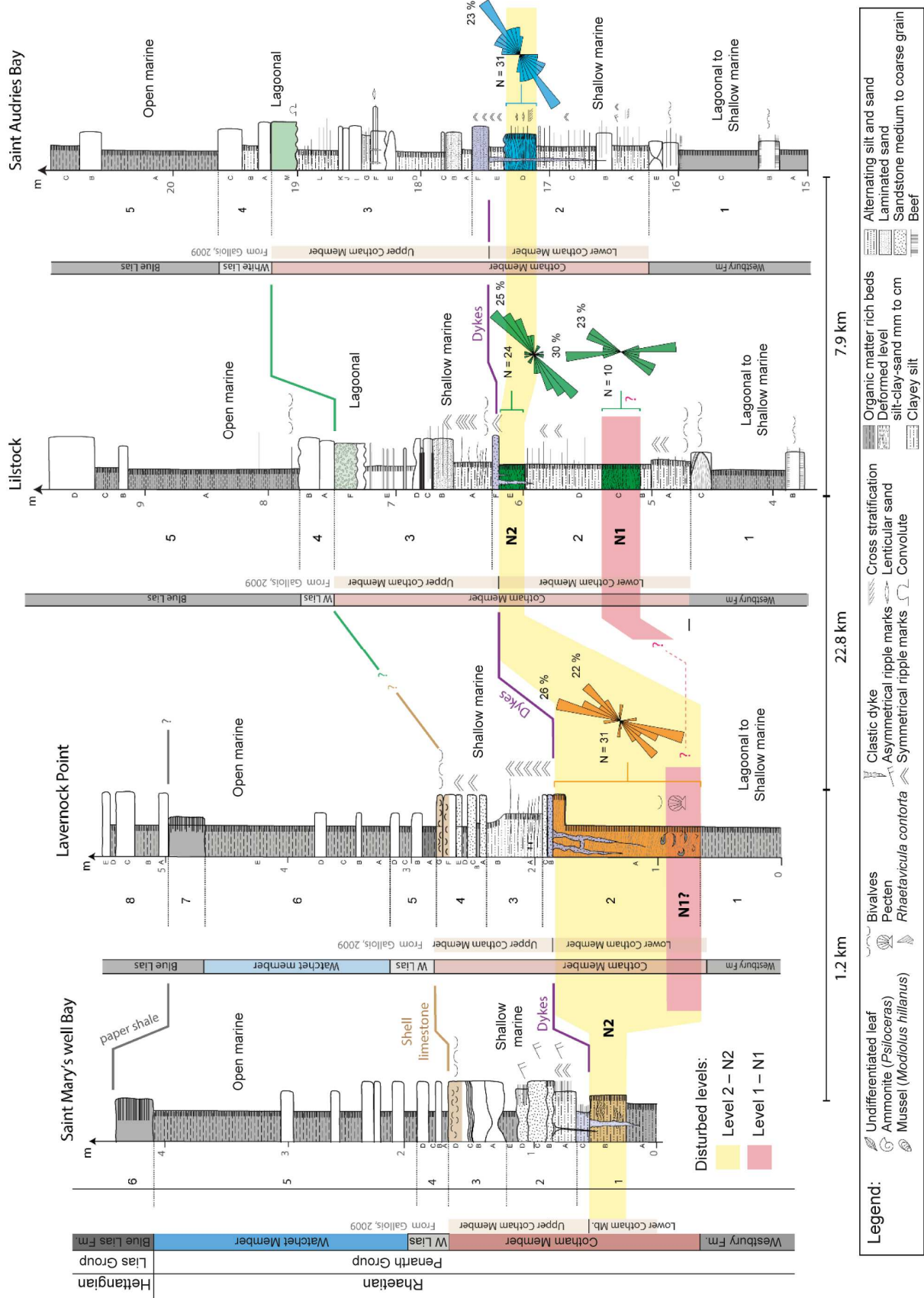
#### 299 4.2.2 Directional characterisation of SSDs

300 Ninety-six fold axes have been measured in the deformed levels of the Cotham Member in the Bristol  
301 Channel area (Fig. 5). Thirty-one folds were measured at each St Audrie's Bay and Lavernock Point  
302 sections, and 34 folds at Lilstock. At St Mary's Well Bay, the observation conditions of the cliff are  
303 insufficient to measure enough fold structures.

304 At St Audrie's Bay, deformations only occur in the 40 cm thick unit 2D, situated just below the layer  
305 that fed the clastic dykes. Despite a moderate dispersion of the data, a preferred NE–SW fold axis  
306 direction, which represent 23% of the data, is recognised in the uppermost part of the lower Cotham  
307 Member. This is in agreement with the results from Northern Ireland, according to which all  
308 deformations in a given level were governed by anisotropic stresses. Accordingly, folds in the 20 cm

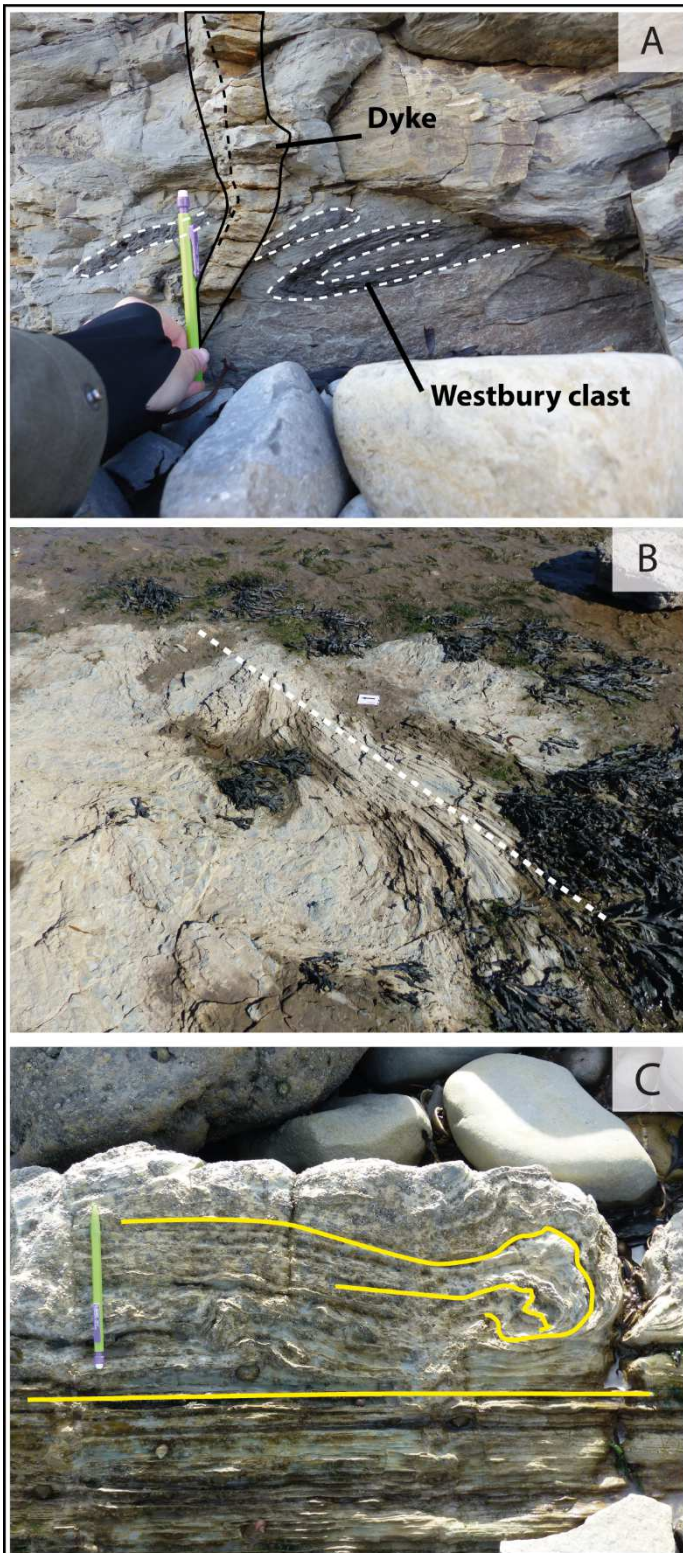
309 thick unit 2E at Lilstock have the same preferred NE–SW trend (25% of the data). Within the  
310 underlying deformed unit 2C, the fold axes show two directions: a main N–S direction (30% of the  
311 data) and a secondary NNE–SSW direction (23% of the data). At Lavernock Point, the fold axes  
312 observed within the deformed beds of the lower Cotham Member which are here 70 cm thick also  
313 show two main directions. A first NE–SW direction (26% of the data, Fig. 5) correspond to the one  
314 observed at St Audrie’s Bay and in unit 2E at Lilstock. A second NNE–SSW direction (22% of the data)  
315 is quite similar to the orientation of the folds within unit 2C at Lilstock.





317 Figure 5. Logged sections and lithostratigraphic correlations within the Bristol Channel area. Rose  
318 diagrams correspond to the main direction axis of folds for each deformed level. Values of the main  
319 peaks are indicated in percentage. N: number of data.

320



321

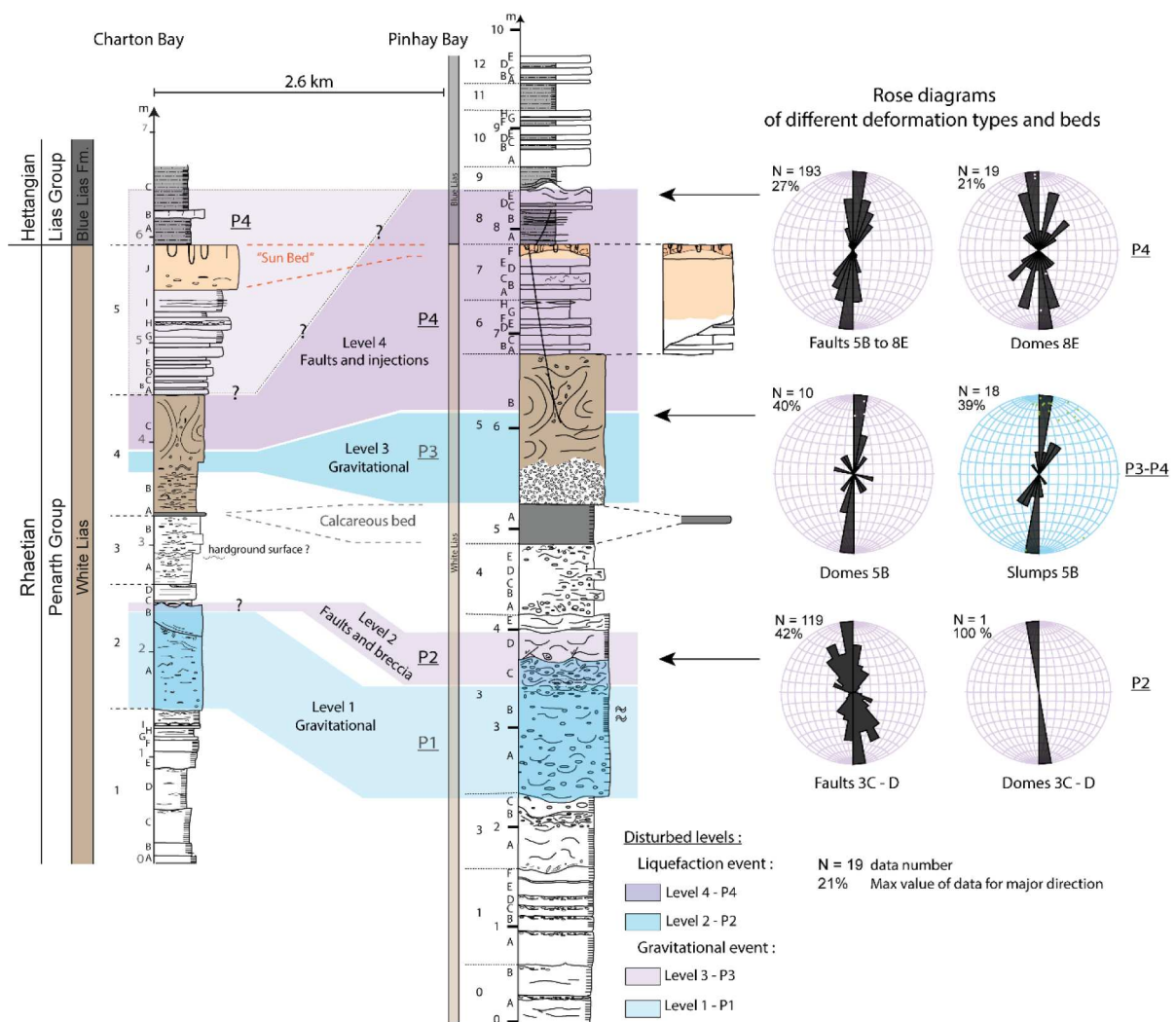
322 Figure 6. SSDs in SW Great Britain (Bristol Channel Basin). (A) Deformed clast of the Westbury  
323 Formation within the base of the lower Cotham Member at Lavernock Point (unit 2A). The clastic  
324 dyke postdates the recumbent fold formation. (B) Isoclinal fold axis in the lower Cotham Member on  
325 the foreshore (unit 2E), Lilstock. (C) Recumbent fold in the lower Cotham Member (unit 2D) at St  
326 Audrie's Bay.

327

328 4.3 South Devon Coast (Wessex Basin)

329 4.3.1 Distribution of SSDs

330 At Charton Bay and Pinhay Bay, South Devon, the “White Lias” facies of the Langport Member is  
 331 more developed than in the Bristol Channel area and corresponds to micritic limestones. It is affected  
 332 by many SSDs of various types and scales. Six sedimentary units and twelve sedimentary units have  
 333 been distinguished at the Charton Bay and Pinhay Bay sections, respectively (Fig. 7). Two major  
 334 deformed levels (levels 2 and 4, in pink color in Fig. 7) have been recognised in both localities.



335

336 Figure 7. Logged sections and position of deformed levels of the South Devon coast area. Rose  
 337 diagrams correspond to the main direction of folds, faults and domes, for each deformed level at  
 338 Pinhay Bay.

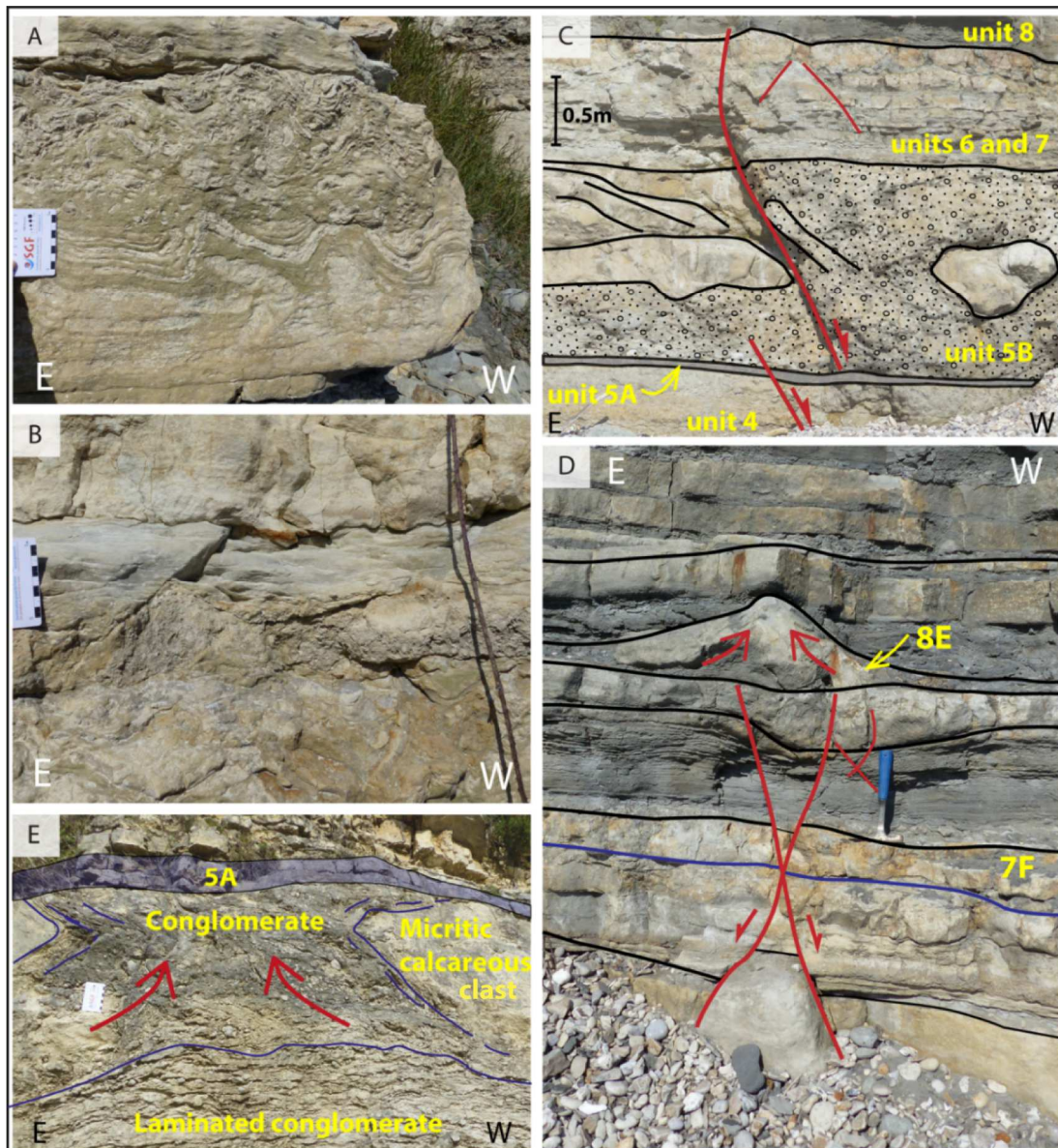
339 Units 2A and 2B at Charton Bay and 3A and 3B at Pinhay Bay, correspond to debris flow deposits.  
340 They are respectively 1.00 m and 1.85 m thick (level 1, Fig. 7) and are composed of a monogenic  
341 calcareous muddy conglomerate with an erosive base and an inverse to undefined grading. At Pinhay  
342 Bay, the upper part of this level is affected by numerous decimetre-scale convolute beds in unit 3B,  
343 which result from important fluid escape features (Fig. 8A).

344  
345 At Pinhay Bay, unit 3C, just above the debris flow deposit, is almost entirely brecciated. The 3C–3D  
346 boundary is affected by small normal faults being about 50 cm in length and having an up to 10  
347 centimetres net slip (level 2, Fig. 7 and Fig. 8B). The syn-sedimentary character of those faults is  
348 suggested by bed thickness variations between the footwall and hangingwall. Within unit 3C,  
349 liquefied material have been molded along the faults as injective domes and peaks. The location and  
350 shape of the injective bodies seem to be clearly controlled by faults. It suggests that the normal faults  
351 guided the upward injections. This deformation required a semi-indurated material to allow brittle  
352 deformation, brecciation and liquefaction of the sediment. No faults have been observed at Charton  
353 Bay, only some rare fluid escape features were observed at the top of unit 2B (Fig. 7).

354  
355 At Pinhay Bay, unit 5B corresponds to a 1.50 m thick disturbed interval. The base of the bed is  
356 completely brecciated and corresponds to an accumulation of angular clasts (level 3, Fig. 7). The  
357 outcrop is also affected by several metre-scale normal faults, which cut units 5B to 8E (level 4, Fig. 7).  
358 To the east, unit 5B is totally disorganised and some metre-scale ball-shaped elements are preserved  
359 in the middle part of the layer. As for units 3C and 3D, the faults that affected unit 5B clearly delimit  
360 the preserved versus liquefied zones. The injective domes are molded along the faults, which appear  
361 to have been at the same time a guide for vertical movement of the liquefied sediment and a  
362 horizontal barrier to deformation (Fig. 8C). In the western part of the section, the faults are less  
363 prominent and the initial deposit of unit 5B is partially preserved. The initial structure of unit 5B  
364 corresponds to a slumped decimetre-scale package of beds. Units 6 and 7 are preserved, but have  
365 faults cutting across them. Unit 8E shows the occurrence of a series of injective domes or bulges with  
366 an upward domed shape. These features are all observed at the top of normal faults (Fig. 8D). They  
367 indicate the local liquefaction and upward flow of the soft sediment material caused by the  
368 displacement along the underlying normal faults. These faults have affected the sedimentary pile up  
369 to unit 8E. The injections have also deformed the overlying bed.

370





371  
 372 Figure 8. SSDs observed in South Devon (A) Fluid escapes features highlighted by convolute bedding  
 373 at the top of a debris flow, unit 3A, Pinhay Bay. (B) Synsedimentary fault between units 3C and 3D at  
 374 Pinhay Bay. Unit 3C is injected along the fault and forms a dome structure. (C) Normal fault affecting  
 375 units 5B to 8E at Pinhay Bay. The fault limits the liquefaction of the slumped unit 5B (shown by  
 376 stippled pattern). (D) Normal fault with injection feature at the top of unit 8E. (E) Massive injection  
 377 features within a matrix-supported conglomerate throughout unit 4C at Charton Bay. Unit 5A is not  
 378 affected (marked by coloured overlay).

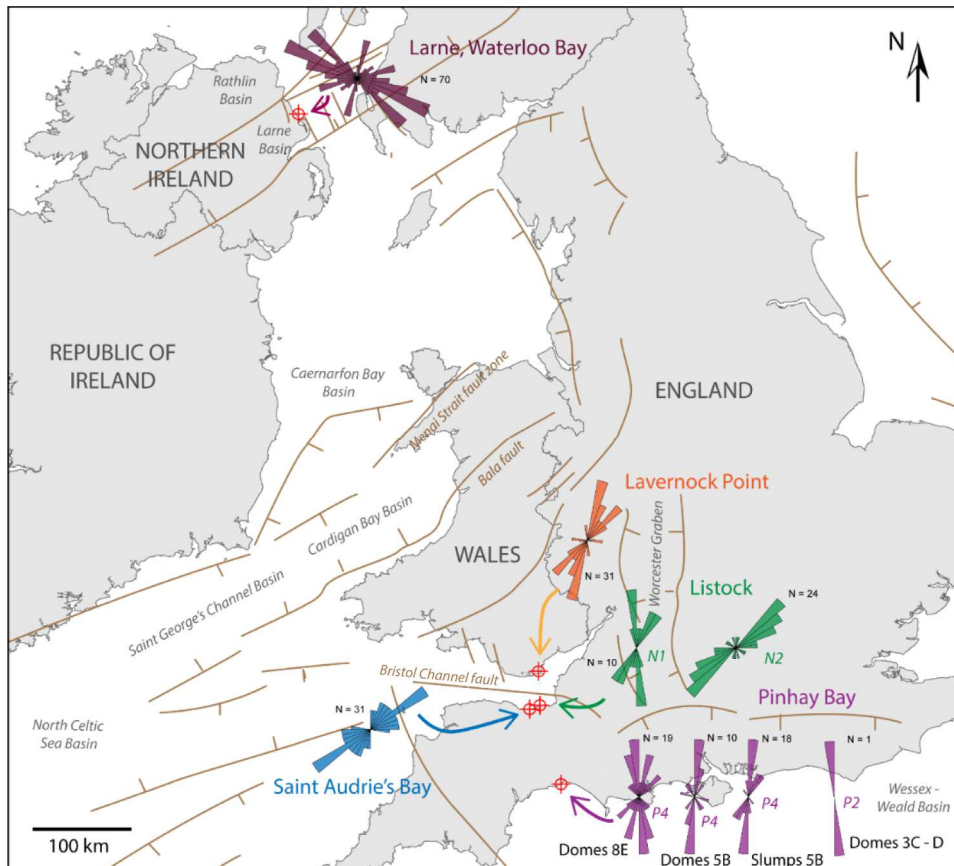
379 At Charton Bay, the unit 4C is a liquefied slumped bed that is similar to unit 5B at Pinhay Bay (level 4,  
 380 Fig. 7). This unit is cut and partially deformed by injections from the underlying beds. It corresponds  
 381 to a micritic limestone affected by numerous fluid escape structures at its base (e.g. convolute  
 382 bedding). The original sediment structure is partially disorganised at the top of the bed. Unlike at  
 383 Pinhay Bay, this interval is not affected by faulting and does not show any dome-like injection  
 384 features directly associated with the faults.

385

386 4.3.2 Directional characterisation of SSDs

387 A statistical study has been carried out on the measurements performed on all types of SSDs axes  
388 from Pinhay Bay. The data include the axis directions of the elongation of injective domes of units 3C  
389 (1 measurement), 5B (10 measurements), 8E (19 measurements), and the axis of the slump in unit 5B  
390 (18 measurements). The orientations of the normal faults affecting units 3C to 3D (119  
391 measurements) and 5E to 8E (193 measurements) were also measured and included in the statistical  
392 analyses (Fig. 7).

393 Both fault sets in units 3C/3D and 5B to 8E have a common N–S to NNE–SSW orientation  
394 (respectively 42% and 27%). These faults show dips towards the east as well as towards the west. The  
395 injective domes in units 3C and 5B, and the helmet-like domes in unit 8E located at the fault tips  
396 show a similar N–S to NNE–SSW direction (respectively 100%, 40% and 21%). Finally, slumps  
397 measured in unit 5B also show a main N–S axis (39%) with a subordinate NNE–SSW orientation. To  
398 summarise, the major directions of structures from each site studied in the UK are shown in the Fig.  
399 9.



400  
 401 Figure 9. Main orientations of deformation structures observed in each outcrop in the UK. N: number  
 402 of data. N1 and N2 in Lilstock, P2 and P4 in Pinhay Bay refer to the two deformed levels recognised in  
 403 these sites.

404

405

406

407

## 408 5. Discussion

### 409 5.1 A seismic origin for Rhaetian SSDs in the UK

410 All observed SSDs occurs in a limited stratigraphic interval at the Triassic–Jurassic boundary in SW  
 411 England, S Wales and Northern Ireland. These SSDs have been observed in four distinct areas over a  
 412 distance of more than 1000 km, namely the Lough Foyle and Larne basins in Northern Ireland, the  
 413 Bristol Channel Basin in the Severn Estuary area (S Wales and N Somerset Coast) and the Wessex  
 414 Basin on the South Devon coast in SW England. **In northern Ireland, in coastal sections to the south of**  
 415 **Waterloo Bay, Jeram et al. (this volume) also described SSDs in the Westbury Fm. at Cloghfin Port,**



416 Islandmagee and in the Lower Cothan Member at Cloghan Point, Whitehead. In addition, SSDs have  
417 been previously documented in other areas of the UK, namely in Central England and North Wales  
418 (Simms, 2007). This spatial distribution suggests a common and large regional extent causal link.

419 Direct or indirect field evidences of earthquake(s) as a trigger of SSDs in the study sites include:

420 (1) All major SSDs and related fluid-escape features being directly molded on and associated with  
421 metric to plurimetric in length normal faults at Pinhay Bay, showing a clear genetic relationship  
422 linking SSDs with faults;

423 (2) Faults axes and dome axes of SSDs measured in the field having a similar N–S to NNE–SSW  
424 direction at Pinhay Bay, again suggesting that SSD distributions were driven by faults activity;

425 (3) The directions of deformation being homogeneous for each disturbed level in all study sites,  
426 suggesting a unique SSD trigger for a given disturbed level;

427 (4) The deformation structures having very well constrained orientations when comparing different  
428 sites (e.g., the N–S to NE–SW directions in south England, Fig. 9), regardless of the deformation type  
429 observed: only earthquake(s) as a trigger may have induced such a (multi-) regional consistency in  
430 SSDs orientations;

431 (5) Numerous laminae thickness variations being observed from either sides of small-scale (up to  
432 several centimetres in length) syn-sedimentary normal micro-faults in Carnduff-1 and Carnduff-2  
433 cores from the Larne Basin, showing the activity of syn-sedimentary micro-faults;

434 (6) The opposite overturning directions of the folds being either oriented towards the north or the  
435 south in the Waterloo Bay foreshore outcrop (Larne Basin) and faults dipping towards the east and  
436 towards the west at Pinhay Bay (Wessex Basin) without preferential orientation. This rules out that  
437 local gravity sliding along slopes is a unique potential trigger of SSD formation;

438 None of the potential other triggers of SSDs such as tsunamis (Takashimizu and Masuda, 2000;  
439 Nanayama et al., 2000; Schnyder et al., 2005; Le Roux et al., 2008 ), tidal flux and tidal bores (Tessier  
440 and Terwindt, 1994; Greb and Archer, 2007), storm waves and breaking waves (Molina et al., 1998)  
441 or rapid sedimentation and loading may explain the direct link observed between SSD and faults  
442 (Pinhay Bay and Larne Basin) and the consistent directions of SSDs at a regional scale in various sites  
443 (e.g., Southern England). As a matter of fact, tidal flux and tidal bores, storm waves, breaking waves  
444 and loading features have no genetic link(s) to fault patterns. They have commonly local effects and  
445 usually do not show inter-regional common directions of structures for various basins, as observed in

446 our study. We do not favor a bolide impact-related tsunami hypothesis to explain the observed SSDs  
447 distribution, because we commonly observed various SSDs axis directions in different stratigraphic  
448 levels. This exclude a unique deformation event, otherwise SSDs should have a unique trend.  
449 Moreover, differences in the fault size in the two deformed levels at Pinhay is also in agreement of  
450 successive shearing events with different intensity. Additional investigations in Northern Ireland by  
451 Jeram et al. (this volume) evidence SSDs occurrence immediately adjacent to faults in the Westbury  
452 Fm at Cloghfin Port and increasing SSDs intensity towards a local fault, directly pointing to a seismic  
453 origin of those SSDs.

454 In addition, the large-scale extensional context in north-western Europe during the Triassic and early  
455 Jurassic (Ziegler and Stampfli, 2001; Ziegler and Dezes, 2006) may certainly have triggered numerous  
456 earthquakes, leading to a context which may have possibly formed earthquakes-induced SSDs.

457 All these observations support a seismic cause as the origin of the liquefaction and the formation of  
458 the disturbed levels. We thus propose that the liquefaction and deformation of the Penarth Group  
459 deposits was due to the activity of local faults during or shortly after sediment deposition, as  
460 suggested by Mayall (1983) and as highlighted by Jeram et al. (this volume) for Northern Ireland  
461 sites.

## 462 *5.2 Role of lithology and fluids in the style of deformation and SSDs localisation*

### 463 *5.2.1 Relation between SSDs type and lithology*

464 Relatively fine-grained sediments like siltstones to fine sandstones and showing a strong vertical  
465 grain size contrast are known to be prone to SSDs formation after suffering an intitial shock,  
466 whatever its origin (Lowe, 1975; Allen, 1982; Montenat et al, 2007). Therefore, in all our study sites,  
467 the fine-grained, alternating lithology, was certainly favourable for SSDs formation. We show that  
468 lithological heterogeneity is also important at various scales, as we observed a role on SSDs  
469 formation for a millimetric to centrimetric lamination of the sedimentary pile (Bristol Channel and  
470 Northern Ireland) and pluri- metric bedding lithological contrasts (Devon).

471 However, differences arise when comparing the style of deformation and its location in the  
472 sedimentary pile in our study sites. In the Bristol Channel and in Northern Ireland, only small normal  
473 faults of centimetre size coexist with recumbent and isoclinal folds of millimetric to metric  
474 dimensions. The original laminations largely remained coherent. On the contrary, in Devon, the  
475 disturbed layers present metre-scale injections and fluid escape features and highly disturbed,  
476 intensely homogenised intervals. Those two different deformations types in the Bristol Channel and

477 in Northern Ireland compared to Pinhay Bay in Devon is probably explained, at least partly, by the  
478 contrasting mechanical properties of the two lithofacies, with a greater carbonate content at Pinhay  
479 Bay than at other sites. Higher carbonate composition at Pinhay Bay may have caused a faster  
480 induration of the depositional mud shortly after deposition, allowing brittle failure of the hardened  
481 sediments after the earthquake shock. The immediately following liquefaction then induced a re-  
482 mobilization of a thick sedimentary pile. On the contrary, in the Bristol Channel and in Northern  
483 Ireland, siliciclastic beds with clay intervals promoted a more ductile behavior and the deformation  
484 was restricted to a rather thin sedimentary pile.

485 In even greater detail, one can observe subtle differences in SSDs style that were probably linked to  
486 lithology. The deformed sediments of Northern Ireland correspond to finely laminated silty-sandy  
487 mudstones, slightly coarser, when compared with the ones deformed around the Bristol Channel  
488 Basin. Furthermore, the thickness of the deformed layer (up to 4.5 m) and the size of SSDs are also  
489 greater in Northern Ireland than in the Bristol Channel Basin. Usual thicknesses of liquefied levels  
490 reported in the literature (Allen, 1977; Plaziat and Ahmamous, 1998; Ken-Tor et al., 2001; Greb and  
491 Archer, 2007) range from a few centimetres to one metre, which highlights the intensity of  
492 deformation in Northern Ireland. The greater susceptibility of silty/sandy materials to liquefaction  
493 (when alternating with more impermeable layers) may be at the origin of the increase in the size of  
494 the folds in Northern Ireland (Obermeier, 1996) and their higher deformed level thicknesses,  
495 **although it has been suggested that the reduced thickness of the SSDs Unit in the Bristol Channel**  
496 **Basin was due to a subsequent erosion, as evidenced by an erosion surface that truncates some of**  
497 **the SSDs (Simms, 2003, 2007).**

498

### 499 *5.2.2 Role of fluids, faulting and overpressure*

500 In the Pinhay Bay section, active faults played a key role in the localisation of deformation in the  
501 beds. As stressed above, the liquefaction and associated SSDs are often localized at the top of faults  
502 (Figs. 7 and 8). On the other hand, the faults also locally played a mechanical barrier role in the  
503 lateral propagation of deformation. As an example, in unit 5B at Pinhay Bay, only the hangingwall of  
504 faults is totally liquefied and de-structured (Fig. 8C), suggesting the occurrence of such a mechanical  
505 barrier role that inhibited the lateral extension of deformation. An additional interesting observation  
506 at Pinhay Bay is that the liquefaction guided by faults only re-mobilise some specific levels, namely  
507 the coarse, brecciated, slumps-rich levels, and not the finer, more homogeneous beds (Fig. 7). The  
508 important incorporation of water during these major gravity events probably facilitated the

509 overpressure of the interstitial water during liquefaction phenomena and led to the brecciation and  
510 then the homogenisation of the initial deposit. Thus, the high porosity and the pre-earthquake(s)  
511 residual water content of these sedimentary units made them very sensitive to the phenomenon of  
512 liquefaction and localized almost all of the deformation. The overlying marl package of the Waterloo  
513 Mudstone Formation in Northern Ireland and the Blue Lias Formation in the Devon may have  
514 enhanced the overpressure processes, helping to concentrate the deformation on the topmost part  
515 of the underlying Penarth Group.

516

### 517 *5.3 Comparing the SSDs and active faults at the Triassic-Jurassic boundary in the UK*

518 One major trend of deformation emerges clearly for each of the study locations, and often  
519 corresponds to the direction of local fault(s) (Fig. 9). In Northern Ireland, the major NW–SE direction  
520 observed at Waterloo Bay, Larne is relatively similar to the trend of Permo-Triassic basin-bounding  
521 faults which have a NNW–SSE trend (Ruffell and Shelton, 1999). The second and minor orientation  
522 found in those sites, well developed at the top of the disturbed level is equivalent to the ENE–WSW  
523 trending Caledonian fault trend (Anderson et al. 1995; Holdsworth et al., 2012). In the Bristol  
524 Channel area, the major N–S or NE–SW direction identified by the SSDs, **in accordance with previous**  
525 **limited results from Simms (2003, 2007)**, is similar to the general orientation of the Cardigan Bay and  
526 Caernarfon basins at the NW of the Bristol Channel (Dobson et al., 1982; Tappin, 1994). These  
527 neighbouring basins are bordered by NE–SW Caledonian faults that were reactivated during the first  
528 opening during Permian and Triassic times, such as the Bala Fault or Menai Straits Fault Zone  
529 (Dobson et al., 1982; Tappin, 1994; Coward, 1995). The N–S trending faults bounding the more distal  
530 Worcester Graben (Fig. 9), may have also been involved in SSDs formation in the Bristol Channel  
531 Basin. The Worcester Basin is bordered by inherited Variscan normal faults (Holdsworth et al., 2012),  
532 but this basin shows a low subsidence during the end of the Triassic (Whittaker, 1985). The known  
533 NE–SW extension direction of the Bristol Channel Basin defined by structural analysis during the  
534 Triassic (Nemcok et al., 1995) is not recorded in the soft sediment deformations, suggesting that the  
535 associated faults, such as the Bristol Channel Fault (Fig. 9), were not responsible for the  
536 earthquake(s) leading to the formation of the observed SSDs. Our record of the fault(s) activity as  
537 shown by SSDs in the Bristol Channel Basin is nevertheless in agreement with the Triassic thickness  
538 maps of the British Isles (Whittaker, 1985). In general, the Bristol Channel Basin experienced very  
539 little subsidence at the end of the Triassic with renewed activity during the Early Jurassic (Whittaker,  
540 1985). In contrast, the Cardigan and Caernarfon basins show thicknesses of Triassic deposits which

541 indicate strong subsidence during this time (Whittaker, 1985), which is consistent with our SSDs  
542 record.

543 The N–S orientation registered in all structures in the Devon does not match with the major N–S  
544 opening direction of the Wessex Basin from the Permian (Hawkes et al., 1998). The single N–S large  
545 regional structure is again the Worcester Graben, but this basin lay at quite a distance toward the  
546 north (Fig. 9). Indeed, the earthquakes that triggered the SSDs in Pinhay may better have occurred on  
547 one of the N-S faults located immediately westward to the Pinhay Bay site, like the Pinhay Fault, the  
548 Rousdon Fault, and/or the Seaton Fault. **In that case, the activity of local faults has presumably  
549 overrided the basinal fault trends influence to produce apparent anomalies in the SSDs trends.**

550

#### 551 *5.4 Implications for the tectonic activity at the Triassic–Jurassic boundary*

552 We have shown that the deformation structures have very well constrained orientations, parallel to  
553 known active faults during the Late Triassic – Early Jurassic. We thus assume that the orientation  
554 defined for each outcrop characterises a local "state of stress" associated with the fault activity. The  
555 SSD patterns may then be used to investigate the active fault or faults system in a given area at the  
556 Triassic–Jurassic boundary.

557

##### 558 *5.4.1 Magnitude of earthquake(s)*

559 To achieve liquefaction in sediments, the magnitude of an earthquake must be greater than  $M_w5$   
560 (Ambraseys and Sarma, 1969; Wang and Manga, 2010). Numerous earthquakes of  $M_w5$  to  $M_w7$  have  
561 already been described in tectonically active areas of extension, such as the Corinth–Patras Rift  
562 (Doutsos and Poulimenos, 1992; Albin et al., 2017) and may have been characteristic of the study  
563 areas in the UK during the Triassic–Jurassic transition. However, most of these earthquakes have  
564 magnitudes of less than  $M_w5$ , as is the case in the African Rift (Lindenfeld et al., 2012) and limits the  
565 number of earthquakes registered by resulting SSDs. Larger earthquakes should result in a more  
566 extensive imprint in the sedimentary archives than moderate ones and according to Obermeier  
567 (1996), a  $M_w7$  earthquake can liquify sediments up to 150 km distance from the epicentre. In the  
568 case of high-magnitude earthquakes ( $>M_w6$ ), the extended duration of the liquefaction phase may  
569 lead to the homogenisation of the alternations, thus forming a mixed unit (Rodríguez-Pascua et al.,  
570 2000). This later state is not observed in the SSDs of Triassic–Jurassic boundary in the UK, apart from  
571 locally one single layer at Pinhay Bay. Thus, the magnitude of the seismic events recorded by SSDs  
572 were probably most of the time not higher than  $M_w5$  to  $M_w6$ .

573

574 5.4.2 A single mega-event or several moderate seismic events over a long duration?

575 The distance between the Waterloo Bay outcrop and the Bristol Channel is over 450 km. Therefore, it  
576 is rather unlikely that a single seismic event may have produced the SSDs in the Lough Foyle, Larne,  
577 Bristol Channel and Wessex basins. Moreover, the thickness of the disturbed layers vary greatly at  
578 each locality, from 0.2 m at Lilstock, to 0.4 m at St Audrie's Bay, and 4.5 metres at Waterloo Bay,  
579 although this has been possibly linked, at least in some localities, to local erosion (Simms, 2007, see  
580 above). The thickness of the de-structured level probably depended on the lithological pattern, as  
581 stressed above, and/or on local erosion. In addition, it may also have been (at least partly) controlled  
582 by the duration of the liquefaction and the magnitude of the earthquake(s), which depend on the  
583 length of the fault (Madariaga and Perrier, 1991). The observed heterogeneity in (1) the thicknesses  
584 of the disturbed layers in various basins and (2) the directional associated pattern as reconstructed  
585 from SSDs support the fact that several earthquakes on local faults rather than one mega-event at the  
586 Triassic–Jurassic transition triggered the SSD patterns. Furthermore, the orientation of the structures  
587 changes in various deformed layers in a single section, as observed in the Bristol Channel Basin.  
588 Jeram et al. (this volume) highlighted the occurrence of two deformation events in the Westbury Fm.  
589 and one deformation event in the Lower Cotham Member in two localities in Northern Ireland, and  
590 interpreted them as seismic-induced. It is therefore most probable that the observed disturbed  
591 layers record a series of different earthquakes, associated with varied active local faults, rather than  
592 a single, mega-event. Indeed, several authors (Mayall, 1983; Hesselbo et al., 2002, 2004; Simms,  
593 2003, 2007) assigned those SSDs to earthquakes but the number of events and their origin are still  
594 debated. Mayall (1983) suggested that the SSDs observed in the UK were linked with extensive  
595 tectonic activity during Triassic and Early Jurassic. More recently, Simms (2003, 2007) suggested that  
596 the cause of the deformation was a rare  $M_w10$  earthquake caused by a meteorite impact, which  
597 remains a possible trigger for the observed SSDs. Hesselbo et al. (2002, 2004) and Lindström et al.  
598 (2015), observed several SSDs in different localities around the Triassic–Jurassic boundary in Western  
599 Tethys that were linked to a succession of earthquakes. They related this geodynamic context with  
600 the development of the magmatic province of the North Atlantic Ocean (CAMP, Central Atlantic  
601 Magmatic Province). Further stressing this relationship, the SSDs found in the Westbury Fm. and in  
602 the Lower Cotham Member in Northern Ireland and in the Bristol Channel Basin immediately  
603 precede an initial carbon isotope excursion correlated with the first major pulse of CAMP volcanism  
604 (Jeram et al., this volume). Our detailed study of the SSDs from the UK favors the idea of a succession  
605 of earthquakes occurring during a very active tectonic period at the Triassic–Jurassic boundary, as  
606 reported by Hesselbo et al. (2002, 2004) and Lindström et al. (2015). The lack of comparable SSDs  
607 anywhere in the Jurassic strata from the UK, whereas evidence of active faulting do occur (Wall and  
608 Jenkyns, 2004) probably relates to the intense geodynamic activity at the Triassic–Jurassic boundary.

609

610 *5.4.3 Rifting pattern at the Triassic–Jurassic boundary during the break-up of Pangea*

611 Our SSDs study shows that there is no principal state of stress common to the entire study region in  
612 the UK. On the contrary, basins of different orientations were active at the same time and were  
613 bordered by faults inherited from the Caledonian and Variscan orogenies, which were successively  
614 re-activated. Some local major faults, potentially active during the Jurassic, do not seem recorded by  
615 the SSDs. As an example, the main normal E–W fault direction, which corresponds to the formation  
616 of the Bristol Channel Basin, recorded in the Triassic deposits of the Bristol Channel area (Nemcok et  
617 al., 1995) is not recorded in the studied disturbed levels. It is possible that the earthquakes  
618 associated with the opening of the Bristol Channel Basin may have been too weak (e.g., lower than  
619  $M_w5$ ) to be recorded, or that the fault was not very active during the deposition of the Penarth  
620 Group to the base of Blue Lias. Conversely, tectonic activity in the Cardigan Bay, St George’s Channel  
621 and North Celtic Sea basins, or on local faults near Pinhay Bay in Devon such as Pinhay Fault, the  
622 Rousdon Fault, and/or the Seaton Fault were probably more intense at this period. The variation of  
623 the directions of SSDs associated with each seismic event finally suggests that the early extension  
624 phases during Pangea Break-up were characterised by the opening of multidirectional, mosaic-style  
625 basins. These basins were established before better constrained and homogeneous extension  
626 directions dominate. Finally, the structural framework inherited from the Caledonian and Variscan  
627 orogenies certainly played an important role in the formation of these Permo-Triassic basins, at least  
628 in these early extension phases.

629

630 **6. Conclusions**

631 The disturbed sedimentary layers observed at the Triassic–Jurassic boundary in the UK and Ireland  
632 combine an array of criteria used to identify seismites, attesting to a seismic origin of the studied  
633 SSDs. Lithology, fluid, faulting, and overpressure processes controlled the style of deformation and its  
634 stratigraphic occurrence. The complete characterisation of the SSDs of the Triassic–Jurassic boundary  
635 strata at the scale of several outcrops across the UK allowed us to associate the directions of SSDs to  
636 the states of stress imposed by the activity of local major faults that were linked to the opening and  
637 development of Triassic basins. A series of seismic events rather than a single, mega seismic shock  
638 was probably responsible for the SSD formations. The variation of the directions of SSDs associated  
639 with each seismic event shows that the early extension phases of the break-up of Pangea were  
640 characterised by the opening of multidirectional basins, largely controlled by the inherited  
641 Caledonian and Variscan tectonic pattern. We can conclude that the orientation of the SSDs in fossil

642 seismites as reconstructed by a detailed structural and field analysis is a useful and promising tool in  
643 the study of past state of stress in sedimentary basins.

644

#### 645 **Acknowledgments**

646 We warmly thank the Geological Survey of Northern Ireland for its welcome and support and  
647 LONMIN (Northern Ireland) for providing full access to the Carnduff-1, Carnduff-2, NIRE 05/08-0002  
648 and NIRE 05/08-0003 cores. We thank Alexis Lethiers (Sorbonne Université) for his help on the field  
649 and for a part of the drawing work. Marine Laborde was funded by a Doctoral Grant from French  
650 authorities hosted at Sorbonne Université and by a SEPM Student Grant. We thank Mike Simms and  
651 an anonymous reviewer for their very helpful and constructive reviews.

652

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