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

Marine Laborde-Casadaban[a], Catherine Homberg[a], Johann Schnyder [a*], Sandra Borderie[b] and Robert Raine[c]

[a]*Laboratoire ISTeP, UMR 7193, case courrier 129, Sorbonne Université, CNRS, 4, pl. Jussieu, 75000, Paris, France.*

[b]*Unit of Earth Sciences, Geosciences Department, University of Fribourg, Chemin du Musée 6, 1700 Fribourg, Switzerland.*

[c]*Geological Survey of Northern Ireland, Dundonald House, Upper Newtownards Road, BT4 3SB, Belfast, Northern Ireland*

*Corresponding author: johann.schnyder@sorbonne-universite.fr

Key Words: Soft sediment deformations, seismites, Rhaetian, Hettangian, Triassic-Jurassic boundary

Abstract

The lagoonal and shallow marine sediments of the Penarth Group in the UK span the Triassic–Jurassic boundary. These sediments contain several disturbed levels with soft sediment deformations (SSDs), such as synsedimentary faults, injective domes, recumbent folds and slumps that are recognised in most basins from SW England and South Wales to NW Northern Ireland. Field observations, notably the closed link of the SSDs to active faults, attest an earthquake origin of the SSDs. Fluids, faults, overpressure and lithology guided the style of the SSDs and their distribution in the sedimentary pile. Analysis of the directional data relating to SSDs in each disturbed level shows preferred orientations of deformation, which correspond to the local state of stress at the time. We favour a series of earthquakes, rather than a single mega-event as a trigger of the observed features. The active local extensional tectonic context was driven by the opening of the Permo-Triassic basins in Western Europe. The data from the SSDs in the UK suggest the development of a multi-directional, mosaic-style extensional context to occur during this early phase of the break-up of Pangea. Our integrated tectonic/sedimentary study suggests that directional data from faults, injective domes, recumbent folds and slumps preserved in sediments are reliable to reconstruct past seismic activity and basin geodynamics.

1. Introduction

Soft sediment deformations (SSDs) are observed in almost all types of environments, from alluvial fan and fluvial systems (Plaziat, 1998); shallow marine and tidal (Pope et al., 1997; Schnyder et al., 2005; Greb and Archer, 2007; Ghosh et al., 2012); lacustrine (Marco et al., 1996; Ken-Tor et al., 2001) and deep marine – turbidite environments (Allen, 1977; Bergerat et al., 2011; Homberg et al., 2013; Basilone et al., 2015). These deformations include convolute bedding, overturned cross-stratification, load structures or water escape features. Due to their low cohesion and small grain size with large water content, silts are particularly sensitive to liquefaction (Allen 1977). During liquefaction, the pore-fluid pressure increases and the grain weights are transferred to the fluid. This causes the loss of grain packing and thus of the internal sediment organisation (Lowe, 1975, 1976; Allen, 1977). The association of liquefaction with a driving force, such as gravity, an unstable density gradient or a shear stress leads to the deformation of the original sedimentary deposit (Owen, 1987). Numerous triggers may cause the liquefaction of sediments such as: (1) tsunamis (Takashimizu and Masuda, 2000; Nanayama et al., 2000; Schnyder et al., 2005; Le Roux et al., 2008), (2) tidal flux and tidal bores (Tessier and Terwindt, 1994; Greb and Archer, 2007), (3) storm waves and breaking waves (Molina et al. , 1998) or (4) rapid sedimentation and loading (Allen, 1982). Earthquake(s) as a trigger of the observed SSDs is frequently proposed in the literature, yet in many cases the liquefaction is attributed to a seismic shock without sufficient evidence (Pope et al., 1997; Kullberg et al., 2001). This shows the difficulty in identifying earthquakes as the cause of sediment liquefaction without well-defined and self-sufficient criteria. Several authors have shown that assigning earthquakes as causing SSDs requires a number of criteria (Wheeler, 2002; Owen and Moretti, 2011), including:

(1) The occurrence of liquefied sediment layers,
(2) A large extent of synchronous liquefied event,
(3) A lateral continuity of a deformed bed along the outcrop,
(4) Similar structures to those observed during recent earthquakes (Audemard and De Santis, 1991; Dechen and Aiping, 2012),
(5) The superposition of SSD levels in the sedimentary pile which may correspond to a succession of seismic events. This criterion is especially strong when it is possible to link the deformed levels to historical earthquakes (Marco et al., 1996; Ken-Tor et al., 2001),
(6) A sedimentary and tectonic context consistent with the occurrence of frequent earthquakes (Bergerat et al., 2011).

Nevertheless, none of these criteria are direct evidence of earthquakes as a trigger. Indeed, very few studies show the direct relationship between syn-sedimentary faults and SSDs (Basilone et al., 2015). The Triassic-Jurassic boundary strata from the UK are affected by numerous soft deformation structures recorded at a large scale (Richardson, 1911; Hallam, 1960; Mayall, 1983; Gallois, 2005, 2007, 2009). Several authors (Mayall, 1983; Hesselbo et al., 2002, 2004; Simms, 2003, 2007) assigned those SSDs to earthquakes but the number of events and their origin are still debated. This study investigates how the orientation of various SSDs are likely to inform on the SSDs trigger in Triassic–Jurassic boundary strata from Southern England, South Wales and Northern Ireland. In the following, we combined sedimentological and structural approaches to investigate the geometry, cross-cutting relationships and distribution of the observed SSDs (faults, folds, slumps, domes), in order to assess the possible earthquake trigger of the disturbed layers and to discuss the relationships of these deformations with the Triassic–Jurassic rifting context.

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

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

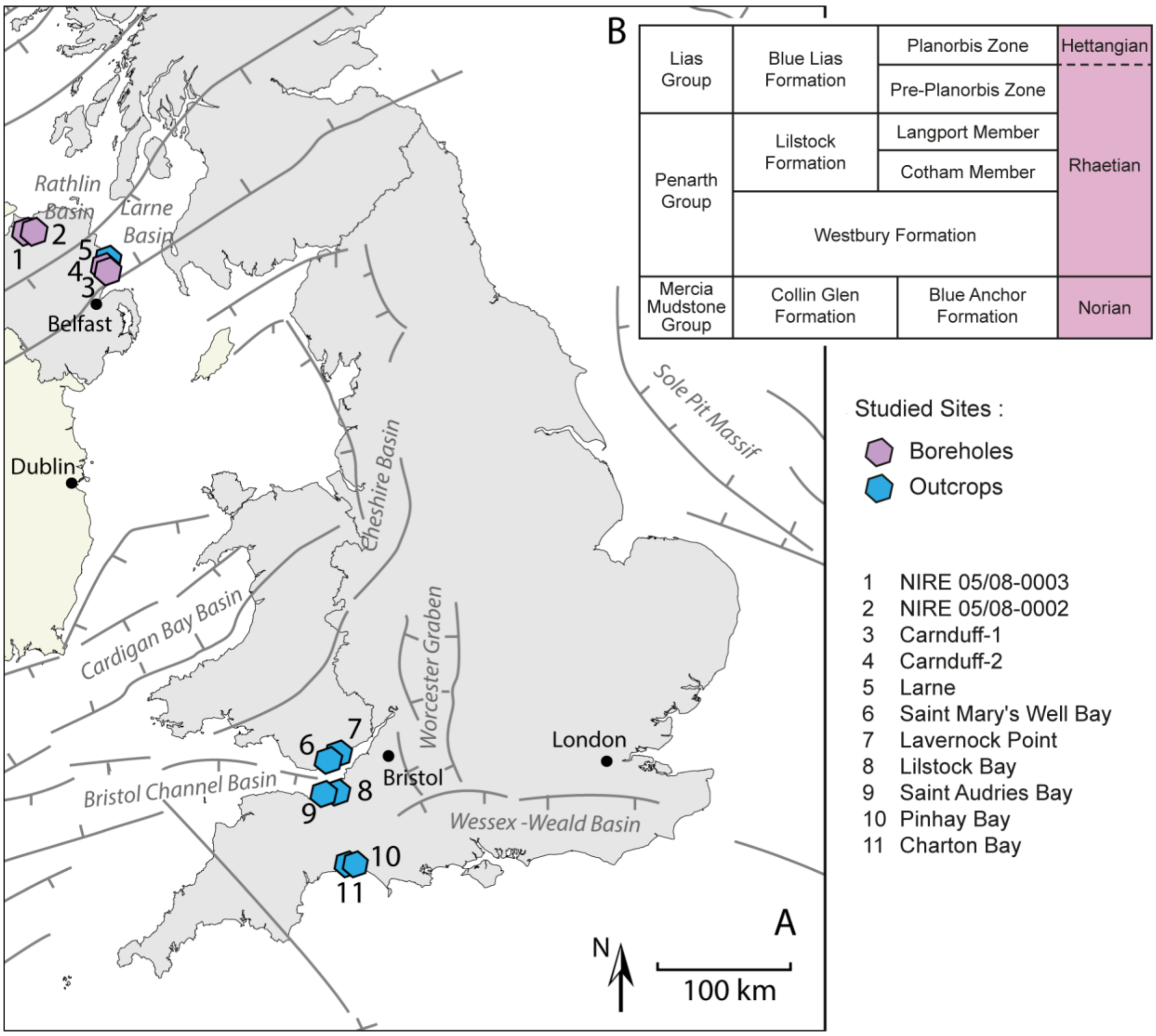


Figure 1. Tectonic framework of the UK at the Triassic–Jurassic transition. A: Location of studied sites. The faults which mark the different Permo-Triassic basins are those defined by Holdsworth et al. (2012). B: Lithostratigraphy of the Triassic-Jurassic boundary deposits in the UK. After Warrington et al. (1994) and Mitchell (2004).

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

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

3. Methods and data

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

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

4. Results: distribution and directional characterisation of SSD in Triassic–Jurassic boundary strata*4.1 Northern Ireland (Larne and Lough Foyle basins)*

4.1.1 Distribution of SSDs

On the Waterloo Bay foreshore section at Larne, the Westbury Formation corresponds to dark grey laminated clays with occasional silty to sandy laminated layers (Figs. 2 and 3). The boundary between the Westbury Formation and the Cotham Member is deformed by centimetre- to decimetre-scale dome structures (Figs. 2 and 3B). This is particularly visible in the western part of the outcrop. The SSDs affect units 1 to 6H (Fig. 2). Dark grey mudstone of the upper Westbury Formation is injected throughout the overlying coarser sediments of the Cotham Member, which pinch out against the dome walls. Claystone and sandstone alternations of the basal Cotham Member are deformed around these injections and form folds caused by the upward movement of the clay. These deformations include diapirs and anticlinal cusps corresponding to fluid escape from overpressured clay up through the non-indurated sandy alternations (Ghosh et al., 2012). The lower Cotham Member is furthermore disturbed by numerous centimetre- to metre-scale isoclinal and recumbent folds (Fig. 3C and D). These folds affect laminated millimetre- to centimetre-thick beds of siltstone, sandstone and claystone (units 1 to 6, Figure 2). The larger folds are in the lenticular unit 2B and have a width and height of 1.0–1.5 m. Second-order folds, with shorter wavelength are observed in the sandier intervals (Figure D). Some areas are stretched and others contracted causing boudinage and variations of sandstone layer thickness, which has amplified the shape of existing wave ripples (Fig. 3D). The interval with SSDs is 4.6 m in thickness. However, the density of SSDs varies laterally, and less deformed areas can be recognised. Unit 3B corresponds to a finer and more homogeneous siltstone that was less affected by soft deformations. From unit 7 to the top of the section, the layers are undisturbed despite having an overall similar lithology to the underlying deformed units (Fig. 2). Only one level with *per-descensum* centimetre-thick dykes was recognised at the top of unit 7. These dykes show multidirectional orientations and were first recorded by Simms and Jeram (2007). They are discussed in detail and interpreted by Jeram et al. (this volume) as two or three generations of subaqueous sedimentary (so-called syneresis) cracking, possibly linked to a seismic event.

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

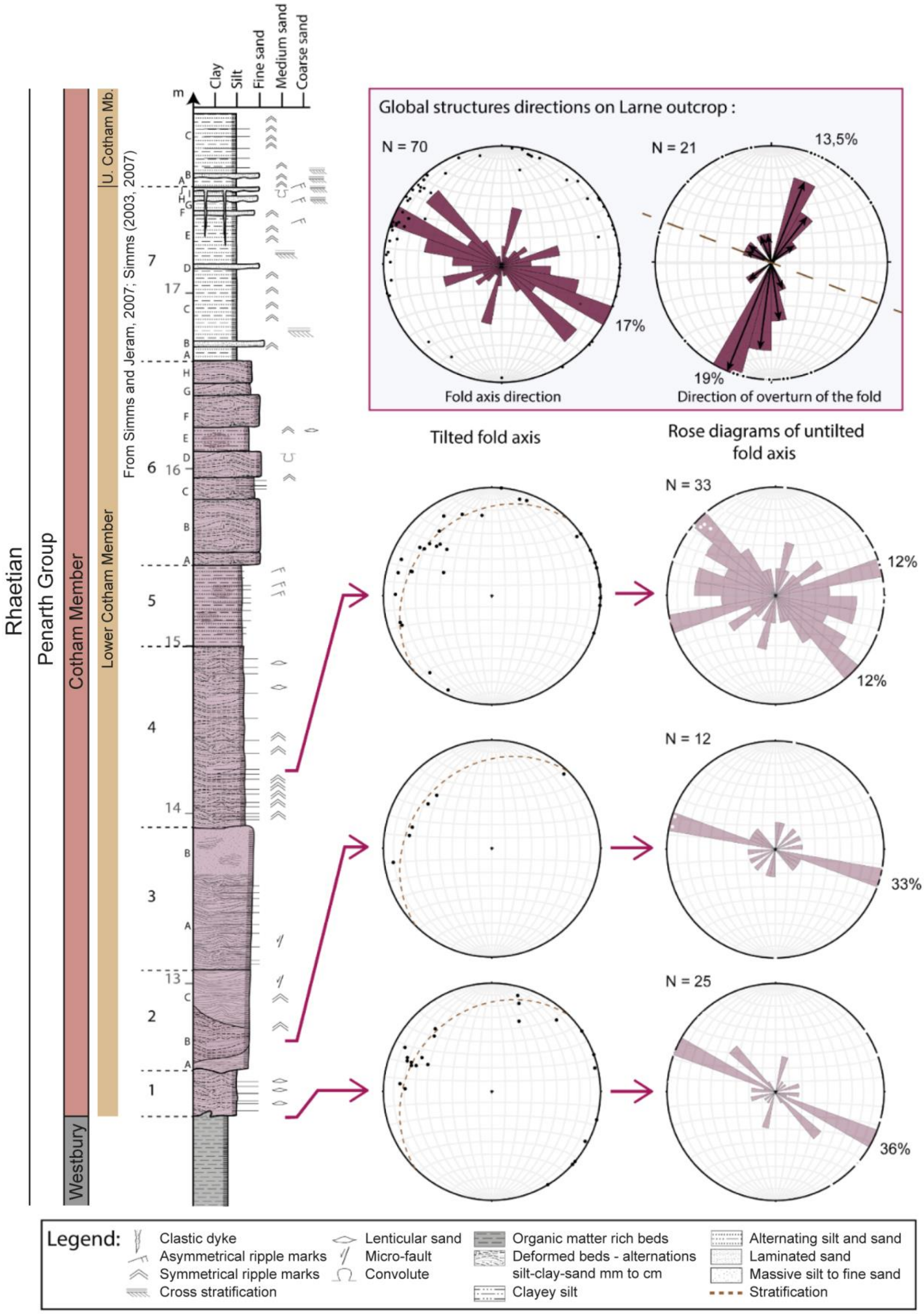


Figure 2. Detailed log section of the Waterloo Bay foreshore, Larne. SSDs are observed in units 1 to 6 (beds in purple).Plots on the left correspond to fold axes with the present day dip. Rose diagrams on

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

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

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

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

Sand- and shell-rich levels disturbed by convolute bedding and synsedimentary faults of millimetre- to centimetre displacement in the Westbury Formation are interpreted as a moderate deformation event (level 1, Figure 4) that affected the Westbury Formation in the Lough Foyle Basin. In our studied sites in the Larne Basin, some convolute bedded shelly levels with burrows are found at the top of the Westbury Formation (Fig. 4), and are likely to equate to this deformation event, as other SSDs have been found in the upper half of the Westbury Fm. in Cloghfin Port, South of Islandmagee (Jeram, et al. this volume). However, the absence of normal faults and of typical liquefaction structures does not rule out the the possibility of a bioturbation origin.

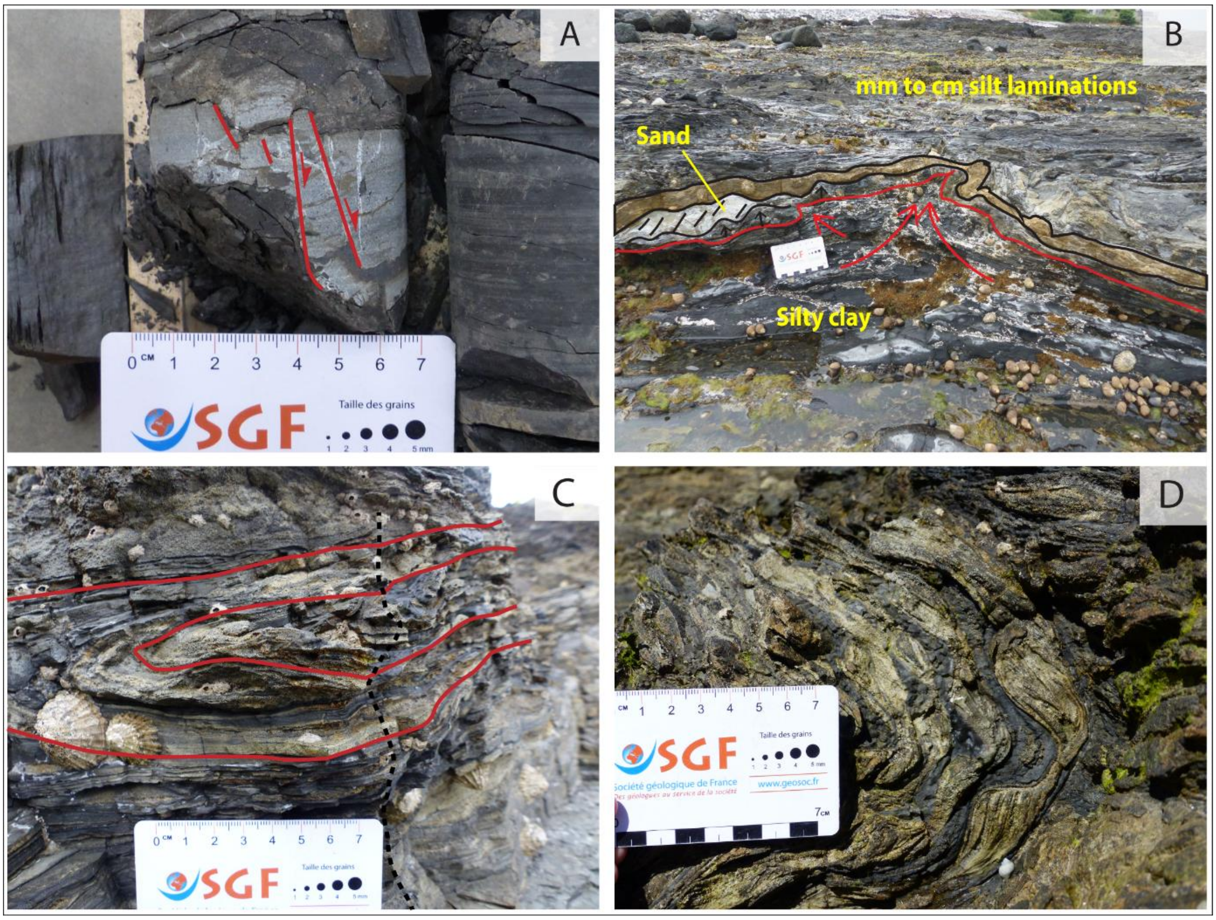


Figure 3. SSDs observed in Northern Ireland. (A) Synsedimentary micro-faults in the Westbury Formation, NIRE 05/08-0003 core. (B) Diapir injection structure at the boundary between the Westbury Formation and the Cotham Member, Waterloo Bay. (C) Recumbent fold in the lower Cotham Member, Waterloo Bay.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

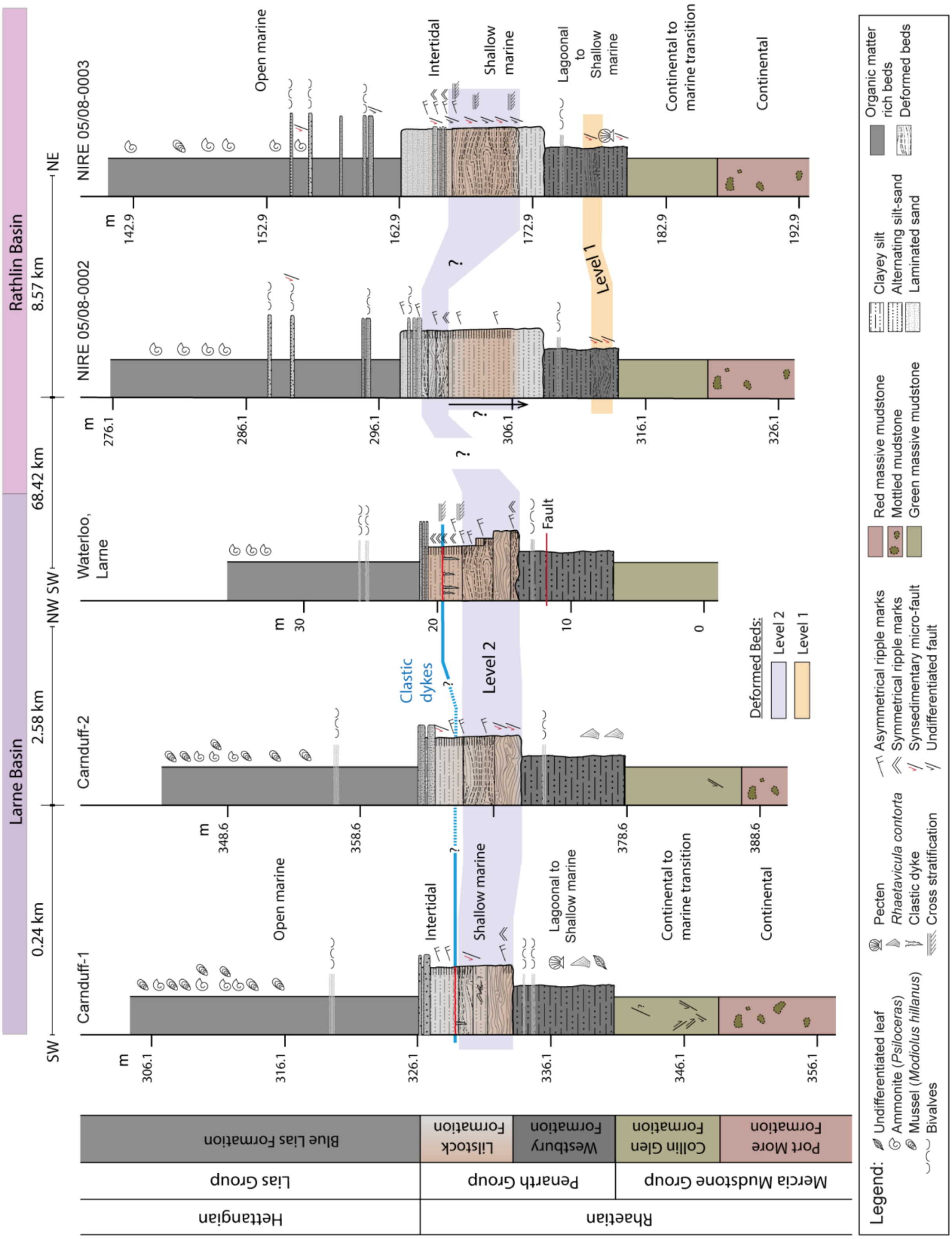


Figure 4**:** Log sections and position of deformed layers in Northern Ireland. Carnduff-1, Carnduff-2, NIRE 05/08-0002 and NIRE 05/08-0003 correspond to cores, and the Waterloo, Larne, corresponds to the outcrop in Waterloo Bay foreshore.

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

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

4.2 Bristol Channel Basin

4.2.1 Distribution of SSDs

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

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

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

4.2.2 Directional characterisation of SSDs

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

At St Audrie's Bay, deformations only occur in the 40 cm thick unit 2D, situated just below the layer that fed the clastic dykes. Despite a moderate dispersion of the data, a preferred NE–SW fold axis direction, which represent 23% of the data, is recognised in the uppermost part of the lower Cotham Member. This is in agreement with the results from Northern Ireland, according to which all deformations in a given level were governed by anisotropic stresses. Accordingly, folds in the 20 cm thick unit 2E at Lilstock have the same preferred NE–SW trend (25% of the data). Within the underlying deformed unit 2C, the fold axes show two directions: a main N–S direction (30% of the data) and a secondary NNE–SSW direction (23% of the data). At Lavernock Point, the fold axes observed within the deformed beds of the lower Cotham Member which are here 70 cm thick also show two main directions. A first NE–SW direction (26% of the data, Fig. 5) correspond to the one observed at St Audrie's Bay and in unit 2E at Lilstock. A second NNE–SSW direction (22% of the data) is quite similar to the orientation of the folds within unit 2C at Lilstock.

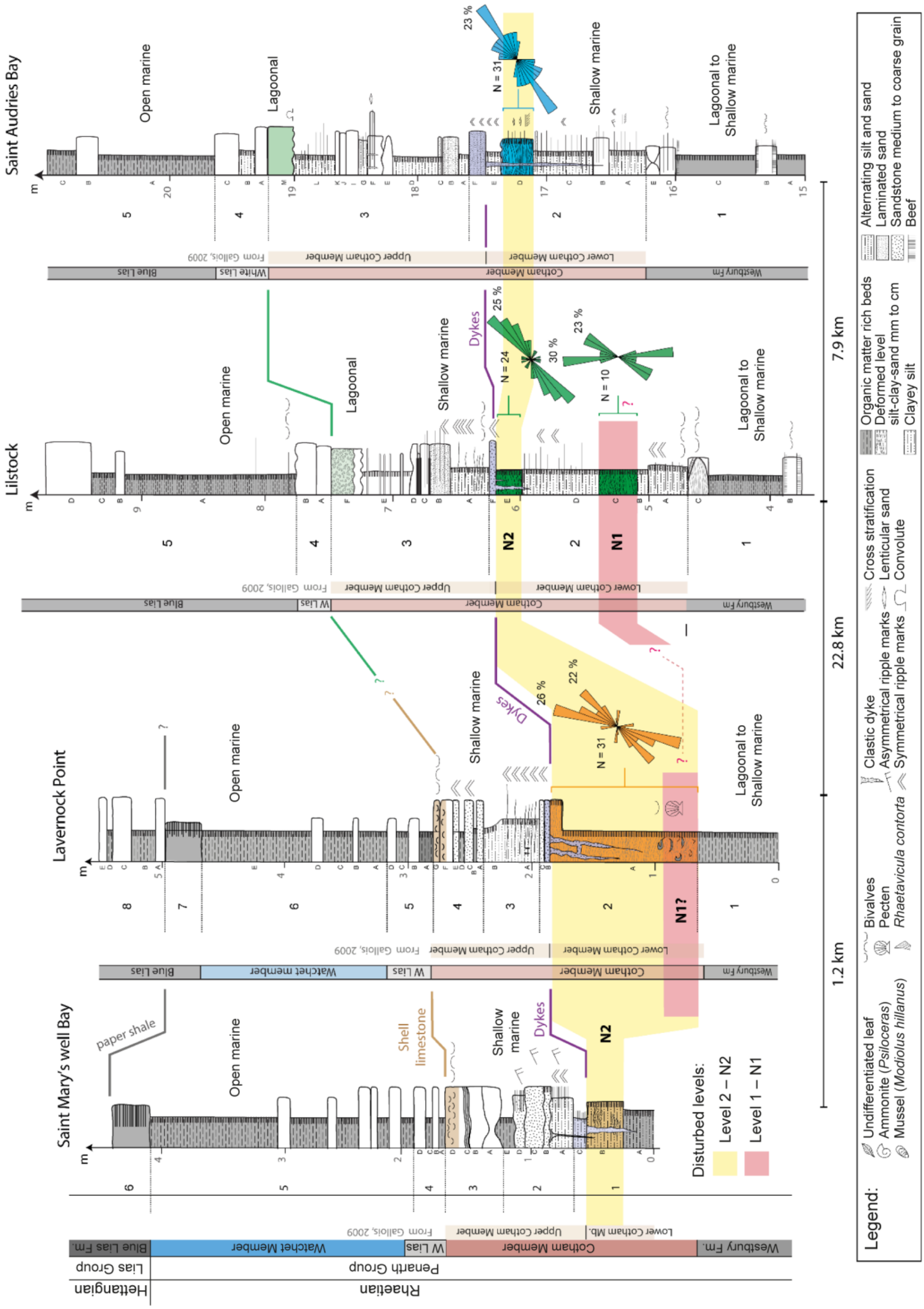

Saint Mary's well Bay
Lavernock Point
Lilstock
Saint Audries Bay
1.2 km
22.8 km
7.9 km
Open marine
Shallow marine
Lagoonal
Lagoonal to Shallow marine
paper shale
Shell limestone
Dykes
N2
N1
N1?
N = 31
N = 24
N = 10
Disturbed levels:
Level 2 – N2
Level 1 – N1
Hettangian
Rhaetian
Lias Group
Penarth Group
Blue Lias Fm
Watchet Member
W Lias
White Lias
Cotham Member
Westbury Fm.
Upper Cotham Member
Lower Cotham Member
From Gallois, 2009
Legend:
Undifferentiated leaf
Ammonite (Psiloceras)
Mussel (Modiolus hillanus)
Bivalves
Pecten
Rhaetavicula contorta
Clastic dyke
Asymmetrical ripple marks
Symmetrical ripple marks
Cross stratification
Lenticular sand
Convolute
Organic matter rich beds
Deformed level
silt-clay-sand mm to cm
Clayey silt
Alternating silt and sand
Laminated sand
Sandstone medium to coarse grain
Beef

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

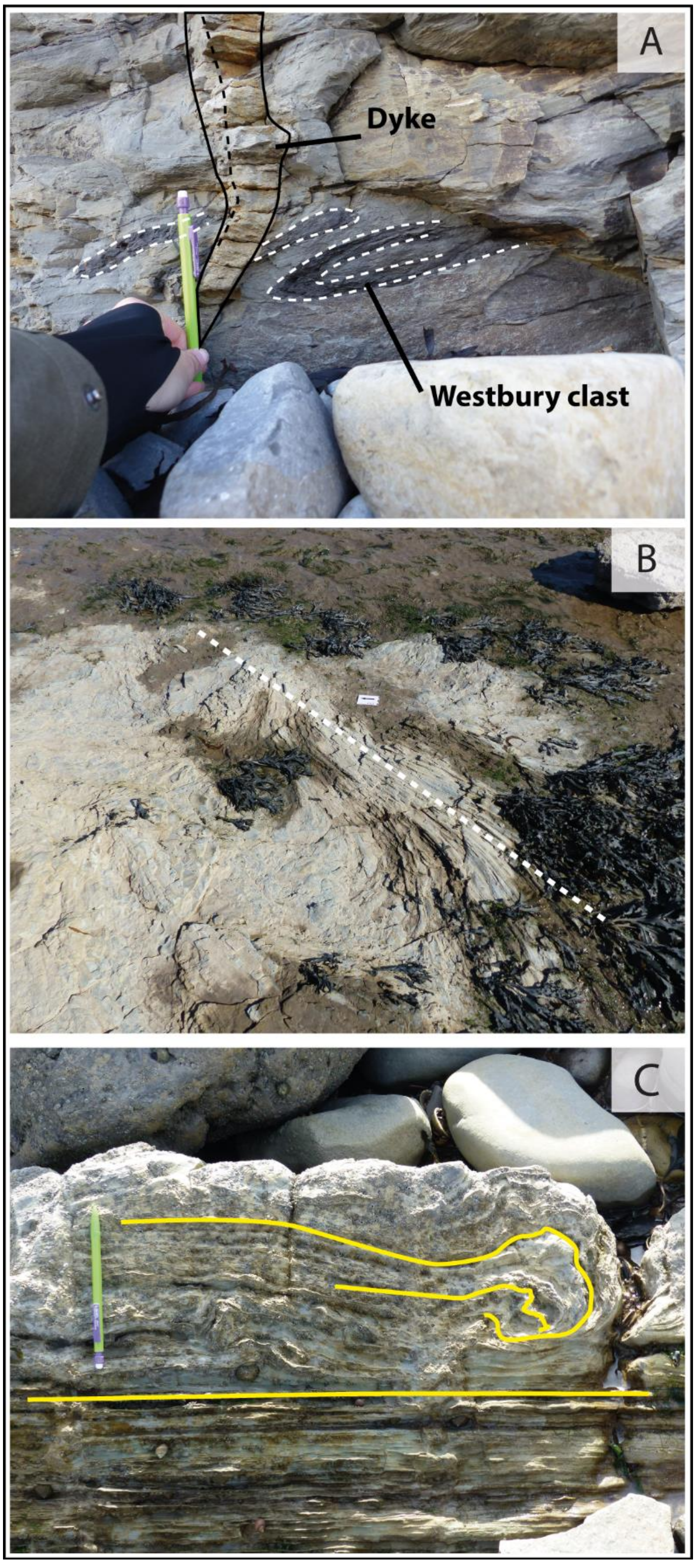


Figure 6. SSDs in SW Great Britain (Bristol Channel Basin). (A) Deformed clast of the Westbury Formation within the base of the lower Cotham Member at Lavernock Point (unit 2A). The clastic dyke postdates the recumbent fold formation. (B) Isoclinal fold axis in the lower Cotham Member on the foreshore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

4.3 South Devon Coast (Wessex Basin)

4.3.1 Distribution of SSDs

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

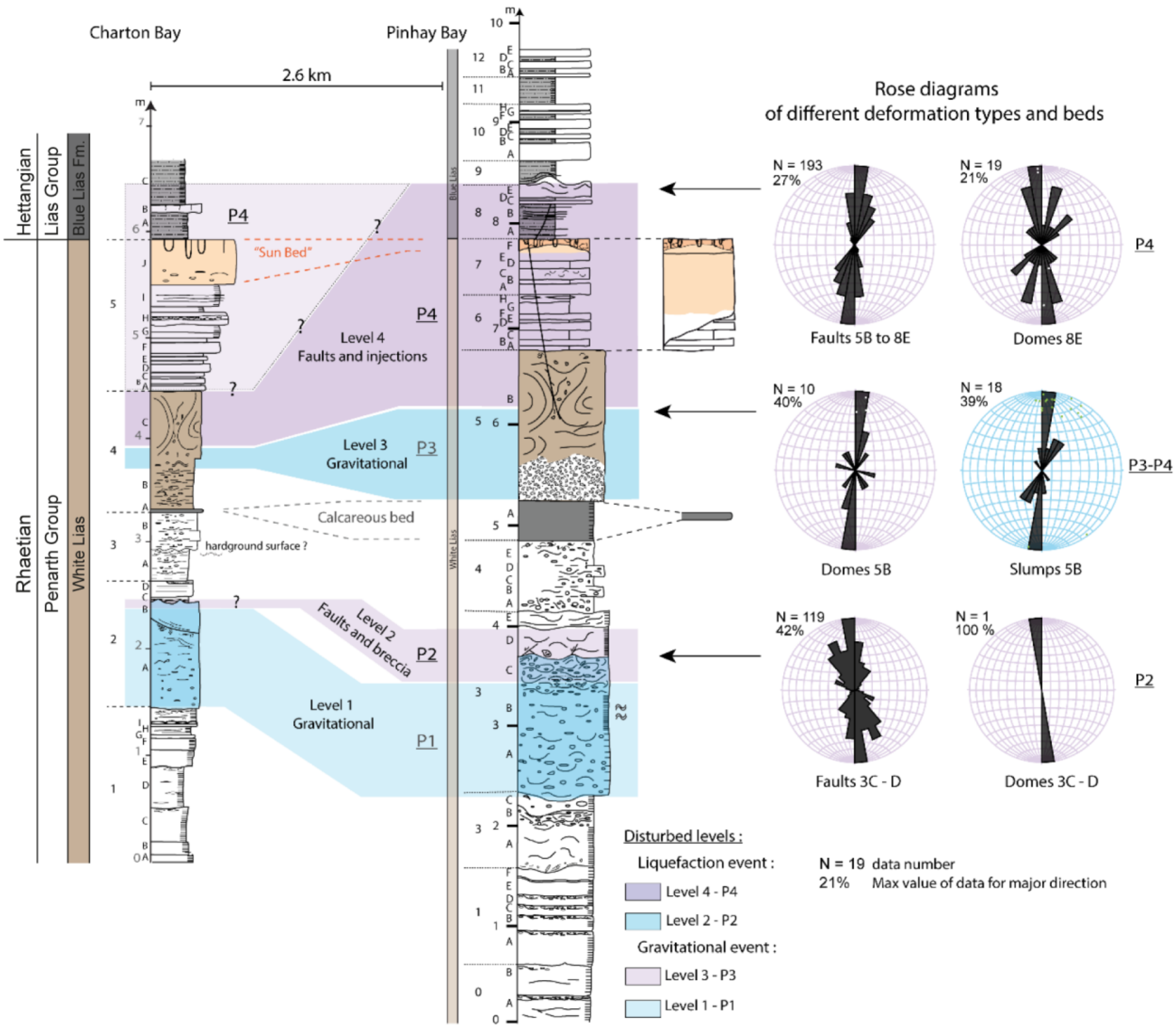


Figure 7. Logged sections and position of deformed levels of the South Devon coast area. Rose diagrams correspond to the main direction of folds, faults and domes, for each deformed level at Pinhay Bay.

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

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

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

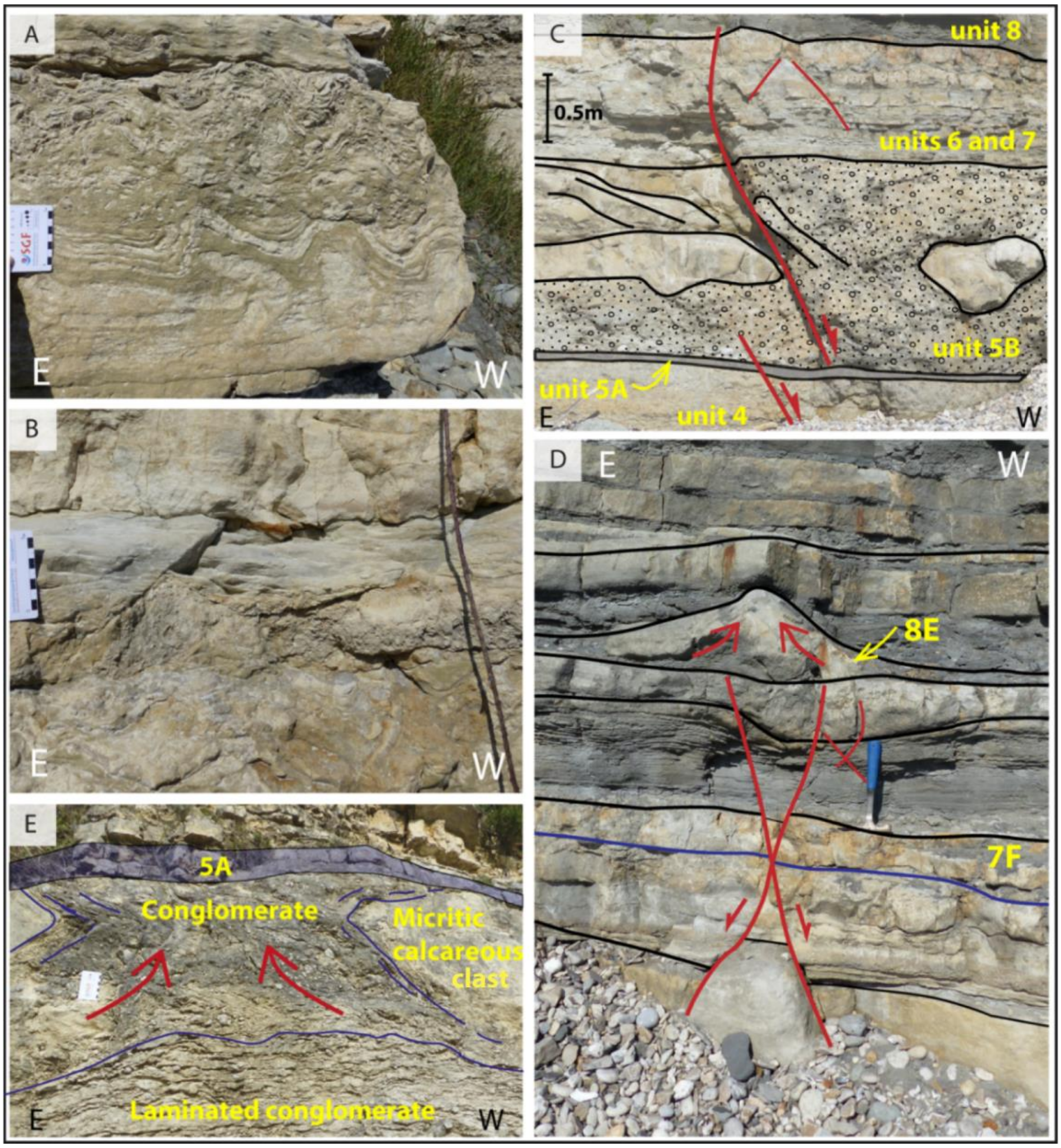


Figure 8. SSDs observed in South Devon (A) Fluid escapes features highlighted by convolute bedding at the top of a debris flow, unit 3A, Pinhay Bay. (B) Synsedimentary fault between units 3C and 3D at Pinhay Bay. Unit 3C is injected along the fault and forms a dome structure. (C) Normal fault affecting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

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

4.3.2 Directional characterisation of SSDs

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

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

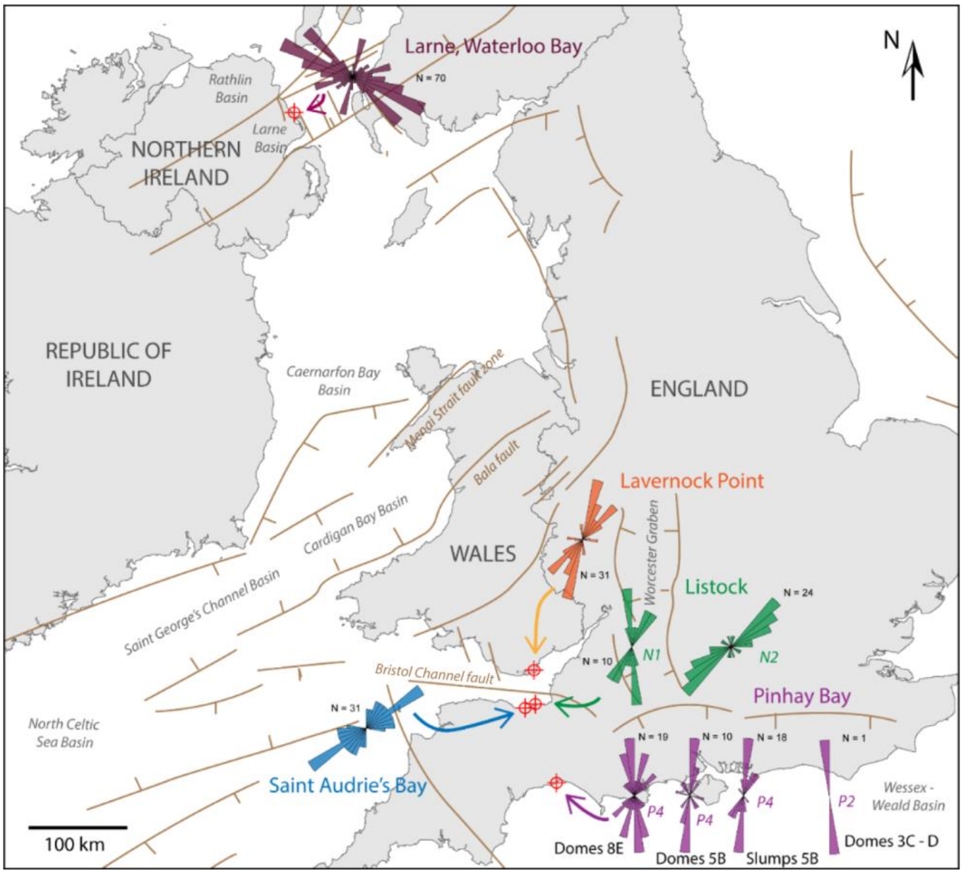


Figure 9. Main orientations of deformation structures observed in each outcrop in the UK. N: number of data. N1 and N2 in Lilstock, P2 and P4 in Pinhay Bay refer to the two deformed levels recognised in these sites.

5. Discussion

5.1 A seismic origin for Rhaetian SSDs in the UK

All observed SSDs occurs in a limited stratigraphic interval at the Triassic–Jurassic boundary in SW England, S Wales and Northern Ireland. These SSDs have been observed in four distinct areas over a distance of more than 1000 km, namely the Lough Foyle and Larne basins in Northern Ireland, the Bristol Channel Basin in the Severn Estuary area (S Wales and N Somerset Coast) and the Wessex Basin on the South Devon coast in SW England. In northern Ireland, in coastal sections to the south of Waterloo Bay, Jeram et al. (this volume) also described SSDs in the Westbury Fm. at Cloghfin Port, Islandmagee and in the Lower Cothan Member at Cloghan Point,Whitehead. In addition, SSDs have been previously documented in other areas of the UK, namely in Central England and North Wales (Simms, 2007). This spatial distribution suggests a common and large regional extent causal link.

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

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

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

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

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

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

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

None of the potential other triggers of SSDs such as tsunamis (Takashimizu and Masuda, 2000; Nanayama et al., 2000; Schnyder et al., 2005; Le Roux et al., 2008), tidal flux and tidal bores (Tessier and Terwindt, 1994; Greb and Archer, 2007), storm waves and breaking waves (Molina et al., 1998) or rapid sedimentation and loading may explain the direct link observed between SSD and faults (Pinhay Bay and Larne Basin) and the consistent directions of SSDs at a regional scale in various sites (e.g., Southern England). As a matter of fact, tidal flux and tidal bores, storm waves, breaking waves and loading features have no genetic link(s) to fault patterns. They have commonly local effects and usually do not show inter-regional common directions of structures for various basins, as observed in our study. We do not favor a bolide impact-related tsunami hypothesis to explain the observed SSDs distribution, because we commonly observed various SSDs axis directions in different stratigraphic levels. This exclude a unique deformation event, otherwise SSDs should have a unique trend. Moreover, differences in the fault size in the two deformed levels at Pinhay is also in agreement of successive shearing events with different intensity. Additional investigations in Northern Ireland by Jeram et al. (this volume) evidence SSDs occurrence immediately adjacent to faults in the Westbury Fm at Cloghfin Port and increasing SSDs intensity towards a local fault, directly pointing to a seismic origin of those SSDs.

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

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

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

5.2.1 Relation between SSDs type and lithology

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

However, differences arise when comparing the style of deformation and its location in the sedimentary pile in our study sites. In the Bristol Channel and in Northern Ireland, only small normal faults of centimetre size coexist with recumbent and isoclinal folds of millimetric to metric dimensions. The original laminations largely remained coherent. On the contrary, in Devon, the disturbed layers present metre-scale injections and fluid escape features and highly disturbed, intensely homogenised intervals. Those two different deformations types in the Bristol Channel and in Northern Ireland compared to Pinhay Bay in Devon is probably explained, at least partly, by the contrasting mechanical properties of the two lithofacies, with a greater carbonate content at Pinhay Bay than at other sites. Higher carbonate composition at Pinhay Bay may have caused a faster induration of the depositional mud shortly after deposition, allowing brittle failure of the hardened sediments after the earthquake shock. The immediately following liquefaction then induced a re-mobilization of a thick sedimentary pile. On the contrary, in the Bristol Channel and in Northern Ireland, siliciclastic beds with clay intervals promoted a more ductile behavior and the deformation was restricted to a rather thin sedimentary pile.

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

5.2.2 Role of fluids, faulting and overpressure

In the Pinhay Bay section, active faults played a key role in the localisation of deformation in the beds. As stressed above, the liquefaction and associated SSDs are often localized at the top of faults (Figs. 7 and 8). On the other hand, the faults also locally played a mechanical barrier role in the lateral propagation of deformation. As an example, in unit 5B at Pinhay Bay, only the hangingwall of faults is totally liquefied and de-structured (Fig. 8C), suggesting the occurrence of such a mechanical barrier role that inhibited the lateral extension of deformation. An additional interesting observation at Pihnay Bay is that the liquefaction guided by faults only re-mobilise some specific levels, namely the coarse, brecciated, slumps-rich levels, and not the finer, more homogeneous beds (Fig. 7). The important incorporation of water during these major gravity events probably facilitated the overpressure of the interstitial water during liquefaction phenomena and led to the brecciation and then the homogenisation of the initial deposit. Thus, the high porosity and the pre-earthquake(s) residual water content of these sedimentary units made them very sensitive to the phenomenon of liquefaction and localized almost all of the deformation. The overlying marl package of the Waterloo Mudstone Formation in Northern Ireland and the Blue Lias Formation in the Devon may have enhanced the overpressure processes, helping to concentrate the deformation on the topmost part of the underlying Penarth Group.

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

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

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

5.4 Implications for the tectonic activity at the Triassic–Jurassic boundary

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

5.4.1 Magnitude of earthquake(s)

To achieve liquefaction in sediments, the magnitude of an earthquake must be greater than M_W5 (Ambraseys and Sarma, 1969; Wang and Manga, 2010). Numerous earthquakes of M_W5 to M_W7 have already been described in tectonically active areas of extension, such as the Corinth–Patras Rift (Doutsos and Poulimenos, 1992; Albini et al., 2017) and may have been characteristic of the study areas in the UK during the Triassic–Jurassic transition. However, most of these earthquakes have magnitudes of less than M_W5, as is the case in the African Rift (Lindenfeld et al., 2012) and limits the number of earthquakes registered by resulting SSDs. Larger earthquakes should result in a more extensive imprint in the sedimentary archives than moderate ones and according to Obermeier (1996), a M_W7 earthquake can liquify sediments up to 150 km distance from the epicentre. In the case of high-magnitude earthquakes (>Mw6), the extended duration of the liquefaction phase may lead to the homogenisation of the alternations, thus forming a mixed unit (Rodríguez-Pascua et al., 2000). This later state is not observed in the SSDs of Triassic–Jurassic boundary in the UK, apart from locally one single layer at Pinhay Bay. Thus, the magnitude of the seismic events recorded by SSDs were probably most of the time not higher than M_W5 to M_W6.

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

The distance between the Waterloo Bay outcrop and the Bristol Channel is over 450 km. Therefore, it is rather unlikely that a single seismic event may have produced the SSDs in the Lough Foyle, Larne, Bristol Channel and Wessex basins. Moreover, the thickness of the disturbed layers vary greatly at each locality, from 0.2 m at Lilstock, to 0.4 m at St Audrie's Bay, and 4.5 metres at Waterloo Bay, although this has been possibly linked, at least in some localities, to local erosion (Simms, 2007, see above). The thickness of the de-structured level probably depended on the lithological pattern, as stressed above, and/or on local erosion. In addition, it may also have been (at least partly) controlled by the duration of the liquefaction and the magnitude of the earthquake(s), which depend on the length of the fault (Madariaga and Perrier, 1991). The observed heterogeneity in (1) the thicknesses of the disturbed layers in various basins and (2) the directional associated pattern as reconstructed from SSDs support the fact that several earthquakes on local faults rather that one mega-event at the Triassic–Jurassic transition triggered the SSD patterns. Furthermore, the orientation of the structures changes in various deformed layers in a single section, as observed in the Bristol Channel Basin. Jeram et al. (this volume) highlighted the occurrence of two deformation events in the Westbury Fm. and one deformation event in the Lower Cotham Member in two localities in Northern Ireland, and interpreted them as seismic-induced. It is therefore most probable that the observed disturbed layers record a series of different earthquakes, associated with varied active local faults, rather than a single, mega-event. Indeed, several authors (Mayall, 1983; Hesselbo et al., 2002, 2004; Simms, 2003, 2007) assigned those SSDs to earthquakes but the number of events and their origin are still debated. Mayall (1983) suggested that the SSDs observed in the UK were linked with extensive tectonic activity during Triassic and Early Jurassic. More recently, Simms (2003, 2007) suggested that the cause of the deformation was a rare M_W10 earthquake caused by a meteorite impact, which remains a possible trigger for the observed SSDs. Hesselbo et al. (2002, 2004) and Lindström et al. (2015), observed several SSDs in different localities around the Triassic–Jurassic boundary in Western Tethys that were linked to a succession of earthquakes. They related this geodynamic context with the development of the magmatic province of the North Atlantic Ocean (CAMP, Central Atlantic Magmatic Province). Further stressing this relationship, the SSDs found in the Westbury Fm. and in the Lower Cotham Member in Northern Ireland and in the Bristol Channel Basin immediately preceed an initial carbon isotope excursion correlated with the first major pulse of CAMP volcanism (Jeram et al., this volume).Our detailled study of the SSDs from the UK favors the idea of a succession of earthquakes occurring during a very active tectonic period at the Triassic-Jurassic boundary, as reported by Hesselbo et al. (2002, 2004) and Lindström et al. (2015). The lack of comparable SSDs anywhere in the Jurassic strata from the UK, whereas evidence of active faulting do occur (Wall and Jenkyns, 2004) probably relates to the intense geodynamic activity at the Triassic-Jurassic boundary.

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

Our SSDs study shows that there is no principal state of stress common to the entire study region in the UK. On the contrary, basins of different orientations were active at the same time and were bordered by faults inherited from the Caledonian and Variscan orogenies, which were successively re-activated. Some local major faults, potentially active during the Jurassic, do not seem recorded by the SSDs. As an example, the main normal E–W fault direction, which corresponds to the formation of the Bristol Channel Basin, recorded in the Triassic deposits of the Bristol Channel area (Nemcok et al., 1995) is not recorded in the studied disturbed levels. It is possible that the earthquakes associated with the opening of the Bristol Channel Basin may have been too weak (e.g., lower than M_W5) to be recorded, or that the fault was not very active during the deposition of the Penarth Group to the base of Blue Lias. Conversely, tectonic activity in the Cardigan Bay, St George's Channel and North Celtic Sea basins, or on local faults near Pinhay Bay in Devon such as Pinhay Fault, the Rousdon Fault, and/or the Seaton Fault were probably more intense at this period. The variation of the directions of SSDs associated with each seismic event finally suggests that the early extension phases during Pangea Break-up were characterised by the opening of multidirectional, mosaic-style basins. These basins were established before better constrained and homogeneous extension directions dominate. Finally, the structural framework inherited from the Caledonian and Variscan orogenies certainly played an important role in the formation of these Permo-Triassic basins, at least in these early extension phases.

6. Conclusions

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

Acknowledgments

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

7. References

Albini, P., Rovida, A., Scotti, O., Lyon-Caen, H., 2017. Large eighteenth–nineteenth century earthquakes in Western Gulf of Corinth with reappraised size and location. Bulletin of the Seismological Society of America 107, 1663–1687. doi:10.1785/0120160181

Allen, J.R.L., (Ed.), 1982. Sedimentary Structures Their Character and Physical Basis Volume 2. Developments in Sedimentology Volume 30B. Elsevier, Amsterdam. pp. 663.

Allen, J.R.L., 1977. The possible mechanics of convolute lamination in gradded sand beds. Journal of the Geological Society, London 134, 19–31.

Ambraseys, N., Sarma, S., 1969. Liquefaction of soils induced by earthquakes. Bulletin of the Seismological Society of America 59, 651–664.

Anderson, T.B., Parnell, J., Ruffell, A.H., 1995. Influence of basement on the geometry of Permo-Triassic basins in the northwest British Isles. In: Boldy, S.A.R. (Ed.), Permian and Triassic Rifting in Northwest Europe. Geological Society, London, Special Publications 91, pp. 103–122.

Audemard, F.A., De Santis, D., 1991. Survey of liquefaction structures induced by recent moderate earthquakes. Bulletin of the International Association of Engineering Geology 44, 5–16.

Basilone, L., Sulli, A., Gasparo Morticelli, M., 2015. The relationships between soft-sediment deformation structures and synsedimentary extensional tectonics in Upper Triassic deep-water carbonate succession (Southern Tethyan rifted continental margin - Central Sicily). Sedimentary Geology 344, 310–322. doi:10.1016/j.sedgeo.2016.01.010

Bazley, R.A.B., Brandon, A., Arthurs, J.W., 1997. Geology of the Country around Limavady and Londonderry. Geological Survey of Northern Ireland, Technical Report. GSNI/97/1. 96 pp.

Bergerat, F., Collin, P.Y., Ganzhorn, A.-C., Baudin, F., Galbrun, B., Rouget, I., Schnyder, J. 2011. Instability structures, synsedimentary faults and turbidites, witnesses of a Liassic seismotectonic activity in the Dauphiné Zone (French Alps): A case example in the Lower Pliensbachian at Saint-Michel-en-Beaumont. Journal of Geodynamics 51, 344–357.

Chadwick, R.A., 1993. Aspects of basin inversion in southern Britain. Journal of the Geological Society, London 150, 311–322.

Coward, M.P., 1995. Structural and tectonic setting of the Permo-Triassic basins of northwest Europe. Geological Society, London, Special Publications 91, 7–39. doi:10.1144/GSL.SP.1995.091.01.02

Dechen, S., Aiping, S., 2012. Typical earthquake-induced soft-sediment deformation structures in the Mesoproterozoic Wumishan Formation, Yongding River Valley, Beijing, China and interpreted earthquake frequency. Journal of Palaeogeography 1, 71–89. doi:http://dx.doi.org/10.3724/SP.J.1261.2012.00007

Deconinck, J.-F., Hesselbo, S.P., Debuisser, N., Averbuch, O., Baudin, F., Bessa, J., 2003. Environmental controls on clay mineralogy of an Early Jurassic mudrock (Blue Lias Formation, southern England). International Journal of Earth Sciences 92, 2, 255–266.

Dobson, M.R., Delanty, L., Whittington, R.J., 1982. Stratigraphy and inversion tectonics in the St. Georges Channel area off SW Wales, U.K. Geo-Marine Letters 2, 23–30.

Doutsos, T., Poulimenos, G., 1992. Geometry and kinematics of active faults and their seismotectonic significance in the western Corinth-Patras rift (Greece). Journal of Structural Geology 14, 689–699. doi:10.1016/0191-8141(92)90126-H

Gallois, R.W., 2005. Correlation of the Triassic and Jurassic succesions proved in the Lyme Regis (1901) Borehole with those exposed on the neaby Devon and Dorset coast. Geoscience in south-west England 11, 99–105.

Gallois, R.W., 2007. Stratigraphy of the of the Penarth Group (Late Triassic) revision of the the East Devon coast. Geoscience in south-west England 11, 287–297.

Gallois, R.W., 2009. Lithostratigraphy of the Penarth Group (Late Triassic) of the Severn Estuary area. Geoscience in south-west England 12, 71–84.

Ghosh, S.K., Pandey, A.K., Pandey, P., Ray, Y., Sinha, S., 2012. Soft-sediment deformation structures from the Paleoproterozoic Damtha Group of Garhwal Lesser Himalaya, India. Sedimentary Geology 261–262, 76–89. doi:10.1016/j.sedgeo.2012.03.006

Greb, S.F., Archer, A.W., 2007. Soft-sediment deformation produced by tides in a meizoseismic area, Turnagain Arm, Alaska. Geology 35, 435–438. doi:10.1130/G23209A.1

Hallam, A., 1960. The White Lias of the Devon coast. Proceedings of the Geologists' Association 71, 47–60. doi:10.1016/S0016-7878(60)80031-4

Hawkes, P.W., Fraser, A.J., Einchcomb, C.C.G., 1998. The tectono-stratigraphic development and exploration history of the Weald and Wessex basins, Southern England, UK. Geological Society, London, Special Publications 133, 39–65. doi:10.1144/GSL.SP.1998.133.01.03

Hesselbo, S.P., Robinson, S.A., Surlyk, F., Piasecki, S., 2002. Terrestrial and marine extinction at the Triassic-Jurassic boundary synchronized with major carbon-cycle perturbation: A link to initiation of massive volcanism? Geology 30, 251. doi:10.1130/0091-7613(2002)030<0251:TAMEAT>2.0.CO;2

Hesselbo, S.P., Robinson, S.A., Surlyk, F., 2004. Sea-level change and facies development across potential Triassic – Jurassic boundary horizons , SW Britain. Journal of the Geological Society, London 161, 365–379.

Holdsworth, R.E., Woodcock, N.H., Strachan, R.A., 2012. Geological Framework of Britain and Ireland. In: Woodcock, N., Strachan, R. (Eds.), Geological History of Britain and Ireland, Second Edition. pp. 19–39, Blackwell Publishing Ltd., Oxford.

Homberg, C., Schnyder, J., Benzaggagh, M. Late Jurassic-Early Cretaceous faulting in the Southeastern French Basin: does it reflect a tectonic reorganization? (2013). Bulletin de la Société Géologique de France, 184, 4-5, p.501-514.

Ivimey-Cook, H.C., 1974. The Permian and Triassic deposits of Wales. In: Owen, T.R. (Ed.), The Upper Paleozoic and Post Paleozoic Rocks of Wales. pp. 295–321. University of Wales Press, Cardiff.

Jeram, A., Simms, M.J., Hesselbo, S.P., Raine, R., 2020. Carbon isotopes, ammonites and earthquakes: Key Triassic-Jurassic boundary events recorded in coastal sections of South East County Antrim, Northern Ireland, UK. Proceedings of the Geologists' Association. This volume.

Ken-Tor, R., Agnon, A., Enzel, Y., Stein, M., Marco, S., Negendank, J.F.W., 2001. High-resolution geological record of historic earthquakes in the Dead Sea basin. Journal of Geophysical Research 106, 2221–2234.

Kullberg, J.C., Olóriz, F., Marques, B., Caetano, P.S., Rocha, R.B., 2001. Flat-pebble conglomerates: A local marker for early jurassic seismicity related to syn-rift tectonics in the Sesimbra area (Lusitanian Basin, Portugal). Sedimentary Geology 139, 49–70. doi:10.1016/S0037-0738(00)00160-3

Le Roux, J.P., Nielsen, S.N., Kemnitz, H., Henriquez, Á., 2008. A Pliocene mega-tsunami deposit and associated features in the Ranquil Formation, southern Chile. Sedimentary Geology 203, 164–180. doi:10.1016/j.sedgeo.2007.12.002

Lindenfeld, M., Rümpker, G., Batte, A., Schumann, A., 2012. Seismicity from February 2006 to September 2007 at the Rwenzori Mountains, East African Rift: Earthquake distribution, magnitudes and source mechanisms. Solid Earth 3, 251–264. doi:10.5194/se-3-251-2012

Lindström, S., Pedersen, G.K., Van De Schootbrugge, B., Hansen, K.H., Kuhlmann, N., Thein, J., Johansson, L., Petersen, H.I., Alwmark, C., Dybkjær, K., Weibel, R., Erlström, M., Nielsen, L.H., Oschmann, W., Tegner, C., 2015. Intense and widespread seismicity during the end-Triassic mass extinction due to emplacement of a large igneous province. Geology 43, 387–390. doi:10.1130/G36444.1

Lowe, D.R., 1976. Subaqueous liquefied and fluidized sediment flows and their deposits. Sedimentology 23, 285–308. doi:10.1111/j.1365-3091.1976.tb00051.x

Lowe, D.R., 1975. Water escape structures in coarse-grained sediments. Sedimentology 22, 157–204. doi:10.1111/j.1365-3091.1975.tb00290.x

Madariaga, R., Perrier, G., 1991. Les tremblements de terres, Presses du CNRS, Paris. pp. 210.

Marco, S., Stein, M., Agnon, A., 1996. Long-term earthquake clustering: A 50,000-year paleoseismic record in the Dead Sea Graben. Journal of Geophysical Research 101, 6179–6191. doi:10.1029/95JB01587

Mayall, M.J., 1983. An earthquake origin for synsedimentary deformation in a late Triassic (Rhaetian) lagoonal sequence , southwest Britain Sedimentary Deformation Desiccation Cracks. Geological Magazine 120, 613–622.

Mitchell, W.I. (Ed.), The Geology of Northern Ireland –Our Natural Foundation (Second Edition), 2004. Geological Survey of Northern Ireland Belfast.

Molina, J.M., Alfaro, P., Moretti, M., Soria, J.M., 1998. Soft-sediment deformation structures induced by cyclic stress of storm waves in tempestites (Miocene, Guadalquivir Basin, Spain). Terra Nova, 10, 145–150.

Montenat, C., Barrier, P., Ott d'Estevou, P., Hibsch, C., 2007. Seismites: An attempt at critical analysis and classification. Sedimentary Geology 196, 1–4, 5–30.

Nanayama, F., Shigeno, K., Satake, K., Shimokawa, K., Koitabashi, S., Miyasaka, S., Ishii, M., 2000. Sedimentary differences between the 1993 Hokkaido-nansei-oki tsunami and the 1959 Miyakojima typhoon at Taisei, southwestern Hokkaido, northern Japan. Sedimentary Geology 135, 255–264. doi:10.1016/S0037-0738(00)00076-2

Nemcok, M., Gayer, R., Miliorizos, M., 1995. Structural analysis of the inverted Bristol Channel Basin: implications for the geometry and timing of fracture porosity. Geological Society, London, Special Publications 88, 355–392. doi:10.1144/GSL.SP.1995.088.01.20

Obermeier, S.F., 1996. Use of liquefaction-induced features for paleoseismic analysis — An overview of how seismic liquefaction features can be distinguished from other features and how their regional distribution and properties of source sediment can be used to infer the location and strength of Holocene paleo-earthquakes. Engineering Geology 44, 1–76. doi:10.1016/S0013-7952(96)00040-3

Owen, G., 1987. Deformation processes in unconsolidated sands. In: Jones, M.E., Prestpn, R.M.F. (Eds.), Deformation of Sediments and Sedimentary Rocks. Geological Society, London, Special Publications 29, pp. 11–24.

Owen, G., Moretti, M., 2011. Identifying triggers for liquefaction-induced soft-sediment deformation in sands. Sedimentary Geology 235, 141–147. doi:10.1016/j.sedgeo.2010.10.003

Plaziat, J.C., Ahmamou, M., 1998. Les différents mécanismes à l'origine de la diversité des séismites , leur identification dans le Pliocène du Saïs de Fès et de Meknès (Maroc) et leur signification tectonique. Geodinamica Acta 11, 183–203. doi:10.1016/S0985-3111(98)80004-1

Pope, M.C., Read, J.F., Hofmann, H.J., Bambach, R., 1997. Late Middle to Late Ordovician seismites of Kentucky, southwest Ohio and Virginia: Sedimentary recorders of earthquakes in the Appalachian basin. GSA Bulletin 109, 489–503. doi:10.1130/0016-7606(1997)109<0489

Radley, J.D., Carpenter, S.C., 1998. The late triassic strata of Manor Farm, Aust, South Gloucestershire, in: Proceedings of the Bristol Naturalists' Society 58, pp. 57–68.

Radley, J.D., Swift, A., 2002. "Sedimentology of the Triassic-Jurassic boundary beds in Pinhay Bay(Devon, SW England)" by P. B. Wignall: comment. Proceedings of the Geological Association 113, 271–272. doi:10.1016/S0016-7878(02)80029-3

Raine, R.J., Copestake, P., Simms, M.J., Boomer, I., 2020. Uppermost Triassic to Lower Jurassic sediments of the island of Ireland and its surrounding basins. Proceedings of the Geologists' Association. This volume.

Richardson, L., 1911. The Rhaetic and contiguous deposits of West, Mid and part of East Somerset. Quarterly Journal of the Geological Society 67, 1–74.

Rodríguez-Pascua, M.A., Calvo, J.P., De Vicente, G., Gómez-Gras, D., 2000. Soft-sediment deformation structures interpreted as seismites in lacustrine sediments of the Prebetic Zone, SE Spain, and their potential use as indicators of earthquake magnitudes during the Late Miocene. Sedimentary Geology 135, 117–135. doi:10.1016/S0037-0738(00)00067-1

Ruffell, A., Shelton, R., 1999. The control of sedimentary facies by climate during phases of crustal extension: Examples from the Triassic of onshore and offshore England and Northern Ireland. Journal of the Geological Society, London 156, 779–789. doi:10.1144/gsjgs.156.4.0779

Schnyder, J., Baudin, F., Deconinck, J.F., 2005. A possible tsunami deposit around the Jurassic-Cretaceous boundary in the Boulonnais area (northern France). Sedimentary Geology 177, 209–227. doi:10.1016/j.sedgeo.2005.02.008

Simms, M.J., 2003. Uniquely extensive seismite from the latest Triassic of the United Kingdom : Evidence for bolide impact ? Geology 31, 557–560. doi:10.1130/0091-7613(2003)031<0557

Simms, M.J., 2007. Uniquely extensive soft-sediment deformation in the Rhaetian of the UK: Evidence for earthquake or impact? Palaeogeography Palaeoclimatology Palaeoecology 244, 407–423. doi:10.1016/j.palaeo.2006.06.037

Simms, M.J., Jeram, A.J., 2007. Waterloo Bay, Larne, Northern Ireland: a candidate Global Stratotype Section and Point for the base of the Hettangian Stage and Jurassic System. ISJS Newsletter 34, 50–68.Stoneley, R., 1982. The structural development of the Wessex Basin. Journal of the Geological Society, London 139, 543–554. doi:10.1144/gsjgs.139.4.0543

Storrs, G., W., 1994. Fossil vertebrate faunas of the British Rhaetian (latest Triassic). Zoological Journal of the Linnean Society, 112, Issue 1-2, 217–259.

Swift, A., 1995. A review of the nature and outcrop of the “White Lias” facies of the Langport Member (Penarth Group: Upper Triassic) in Britain. Proceedings of the Geologists' Association 106, 247–258. doi:10.1016/S0016-7878(08)80236-2

Swift, A., Martill, D.M., (Eds.), 1999. Fossils of the Rhetian Penarth Group. Palaeontological Association, London.

Takashimizu, Y., Masuda, F., 2000. Depositional facies and sedimentary successions of earthquake-induced tsunami deposits in Upper Pleistocene incised valley fills, central Japan. Sedimentary Geology 135, 231–239. doi:10.1016/S0037-0738(00)00074-9

Tappin, D.R., Chadwick, R.A., Jackson, A.A., Wingfield, R.T.R., Smith, N.J.P, 1994. The geology of Cardigan Bay and the Bristol Channel. HMSO, London, pp. 107.

Tessier, B. and Terwindt, J., 1994. An Example of Soft-Sediment Deformations in an Intertidal Environment - The Effect of a Tidal Bore. Comptes-Rendus de l'Académie des Sciences Série II, 217-233.

Wall, G.R.T. and Jenkyns, H.C., 2004. The age, origin and tectonic significance of Mesozoic sediment-filled fissures in the Mendip Hills (SW England): implications for extension models and Jurassic sea-level curve. Geological Magazine, 141, 471-504.

Wang, C.-Y., Manga, M., 2010. Hydrologic responses to earthquakes and a general metric. Geofluids 206–216. doi:10.1111/j.1468-8123.2009.00270.x

Warrington, G., Audley, M.G., Elliotr, R.E., 1980. A correlation of Triassic Rocks in the British Isles. Geological Society, London, Special Report 13.

Warrington, G., Cope, J.C.W., Ivimey-Cook, H.C., 1994. St Audrie’s Bay, Somerset, England: a candidate Global Stratotype Section and Point for the base of the Jurassic System. Geological Magazine 131, 191–200. doi:10.1017/S0016756800010724

Waters, R.A., Lawrence, D.J.D. 1987. Geology of the South Wales Coalfield, Part III, the country around Cardiff. Memoir for 1:50 000 geological sheet 263 (England and Wales). Third Edition. HMSO, London. pp. 114.

Wheeler, R.L., 2002. Distinguishing seismic from nonseismic soft-sediment structures: Criteria from seismic-hazard analysis. In: Ettensohn, F.R., Rast, N., Brett, C.E. (Eds.), Ancient Seismites. Geological Society of America, Special Papers 359. GSA, Boulder, Colorado, pp. 1–11.

Whittaker, A., 1985. Isopachytes for the Mercia Mudstone Group plus Penarth Group. In: Whittaker, A. (Ed.), Atlas of Onshore Sedimentary Basins in England and Wales. Blackie & Son Limited, Glasgow, pp. 80.

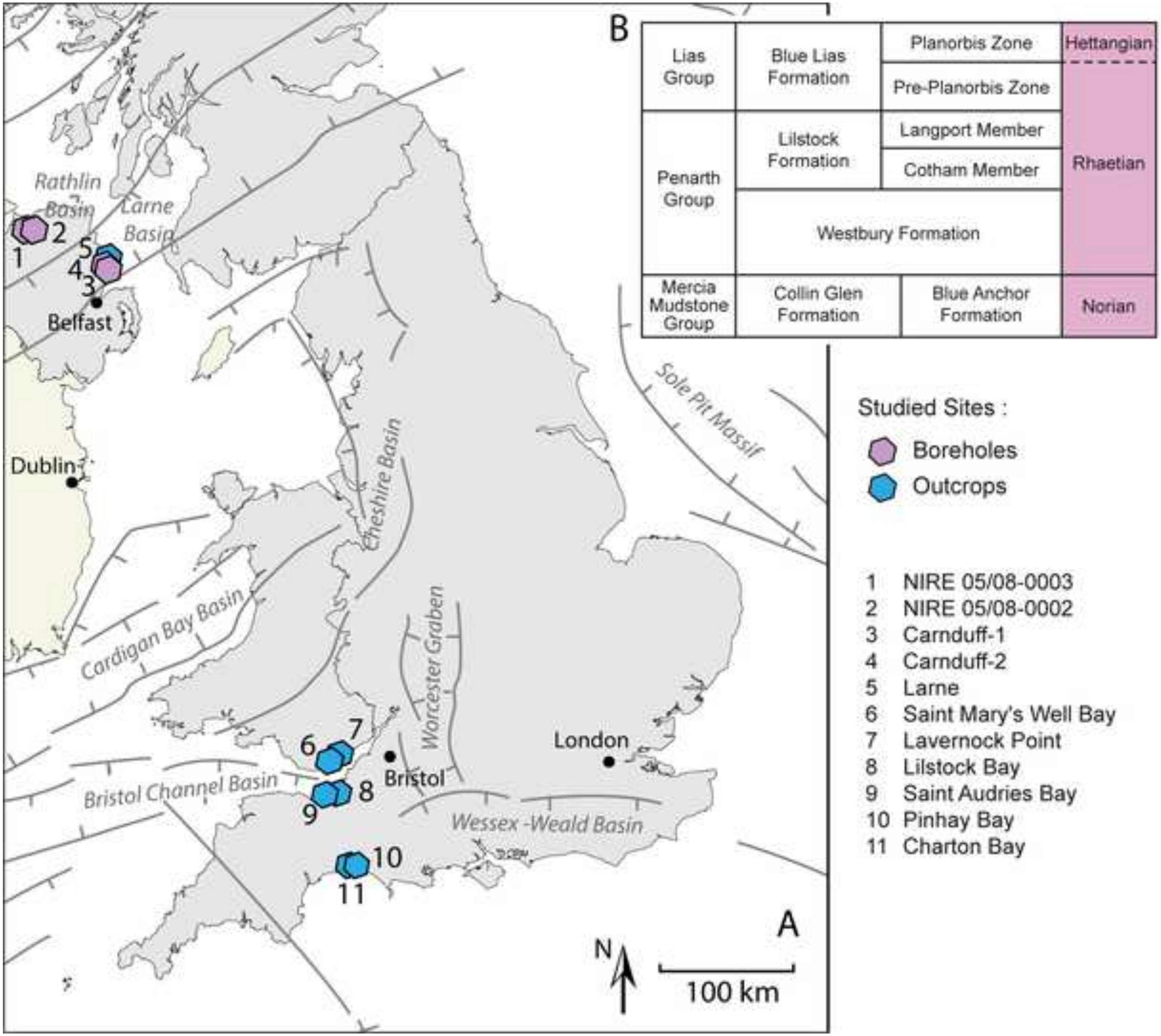
Wignall, P.B. 2001 Sedimentology of the Triassic-Jurassic boundary beds in Pinhay Bay (Devon, SW England). Proceedings of the Geologists' Association 112, 349–360.

Ziegler, P.A., Stampfli, G.M., 2001. Late Paleozoic-Early Mezozoic plate boundary reorganisation: collapse of the variscan orogen and opening of Neothetys., in: Cassinis, G. (Ed.), Permian Continental Deposits of Europe and Other areas.Regional Reports and Correlations. pp. 17–34.

Ziegler, P. a., Dezes, P., 2006. Crustal evolution of Western and Central Europe. Geological Society, London, Memoirs 32, 43–56. doi:10.1144/GSL.MEM.2006.032.01.03

Figure 1

Click here to access/download;Figure;Fig 1.tif



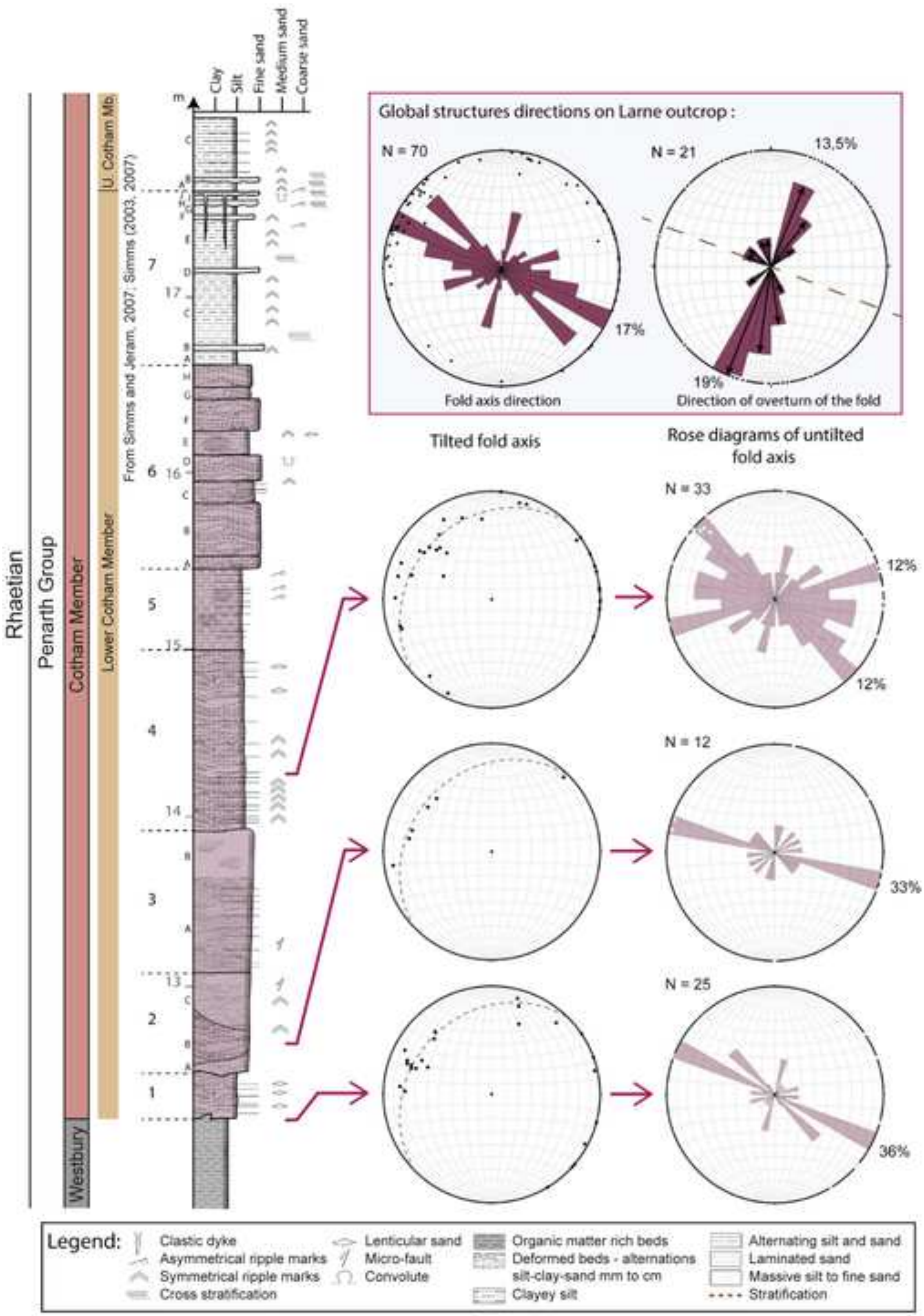

Rhaetian
Penarth Group
Cotham Member
Westbury
U. Cotham Mb
Lower Cotham Member
From Simms and Jeram, 2007; Simms (2003, 2007)
m
Clay
Silt
Fine sand
Medium sand
Coarse sand
Global structures directions on Larne outcrop :
N = 70
17%
Fold axis direction
N = 21
13,5%
19%
Direction of overturn of the fold
Tilted fold axis
Rose diagrams of untilted fold axis
N = 33
12%
12%
N = 12
33%
N = 25
36%
Legend:
Clastic dyke
Asymmetrical ripple marks
Symmetrical ripple marks
Cross stratification
Lenticular sand
Micro-fault
Convolute
Organic matter rich beds
Deformed beds - alternations silt-clay-sand mm to cm
Clayey silt
Alternating silt and sand
Laminated sand
Massive silt to fine sand
Stratification

Figure 3

Click here to access/download;Figure;Fig 3.tif

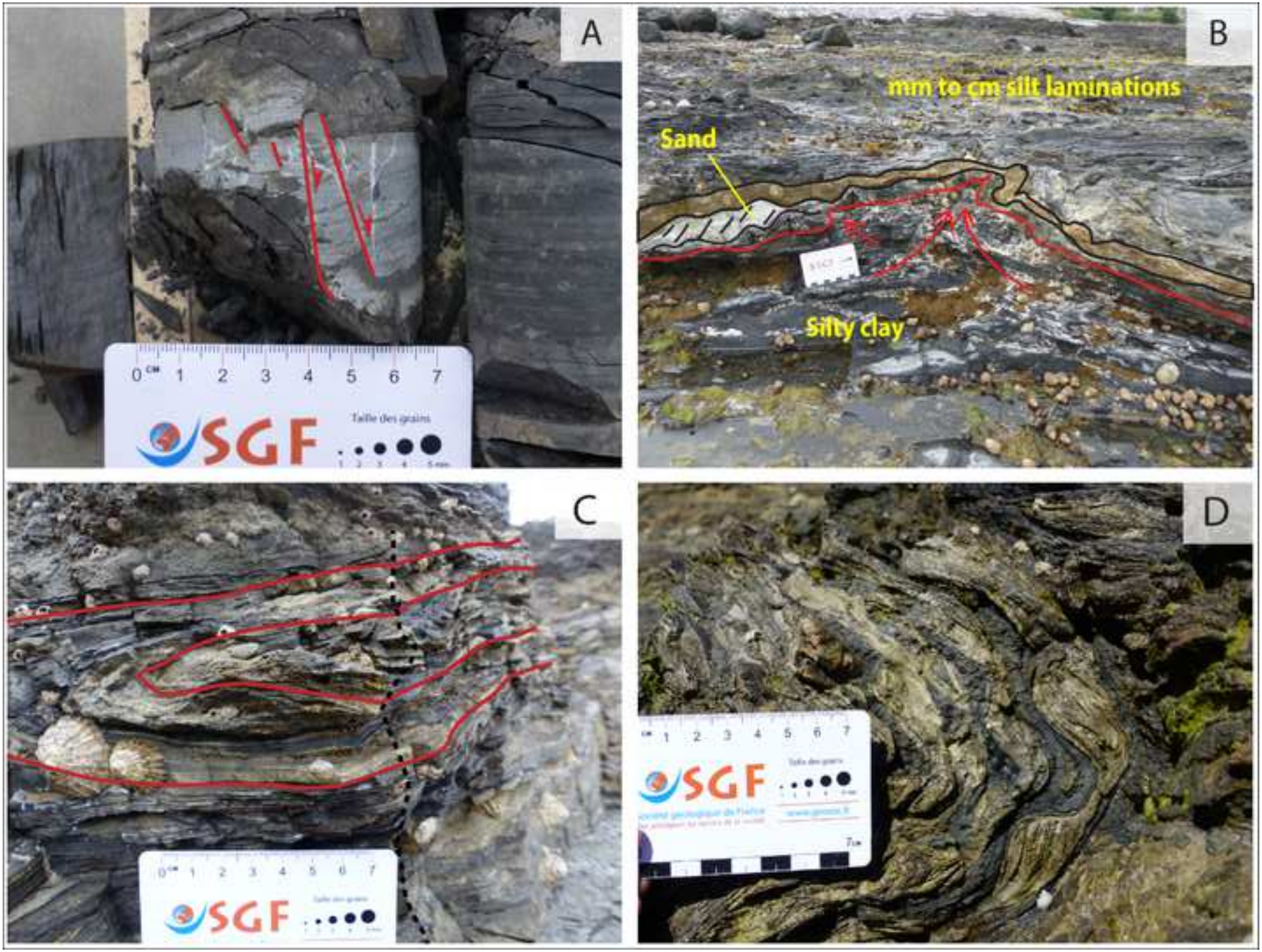


Figure 4

Click here to access/download;Figure;Fig 4.tif

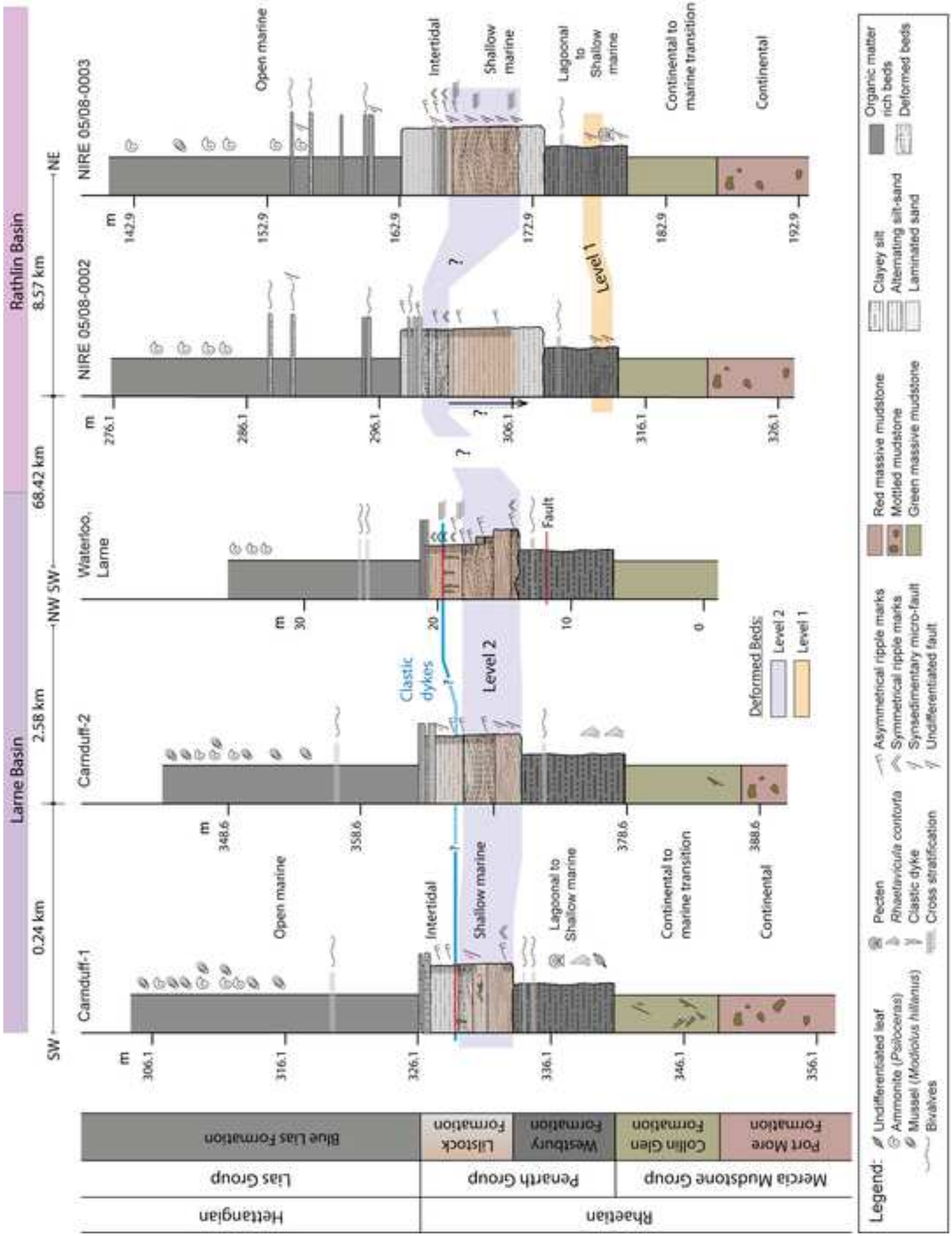


Figure 5

Click here to access/download;Figure;Fig 5.tif

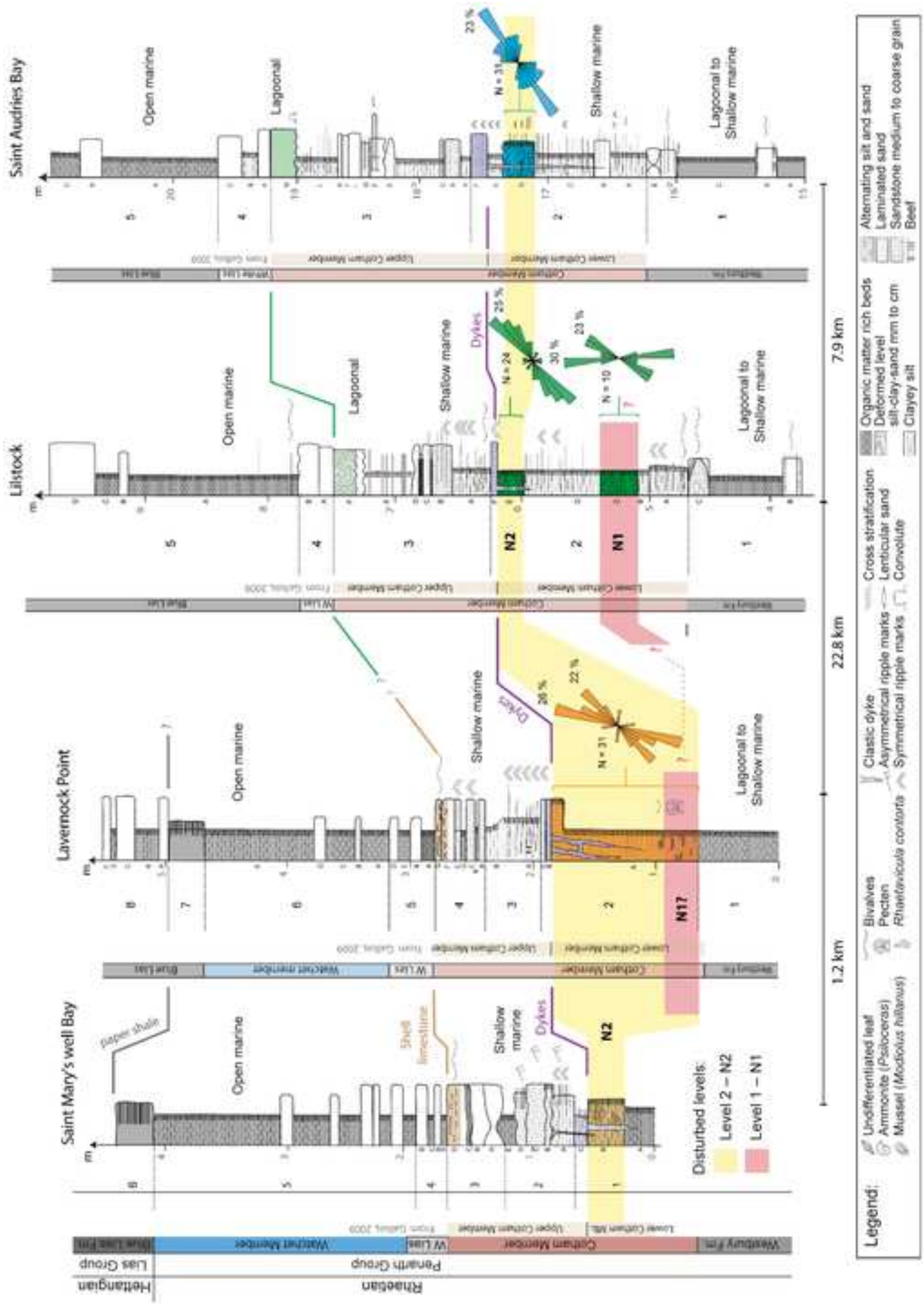


Figure 6

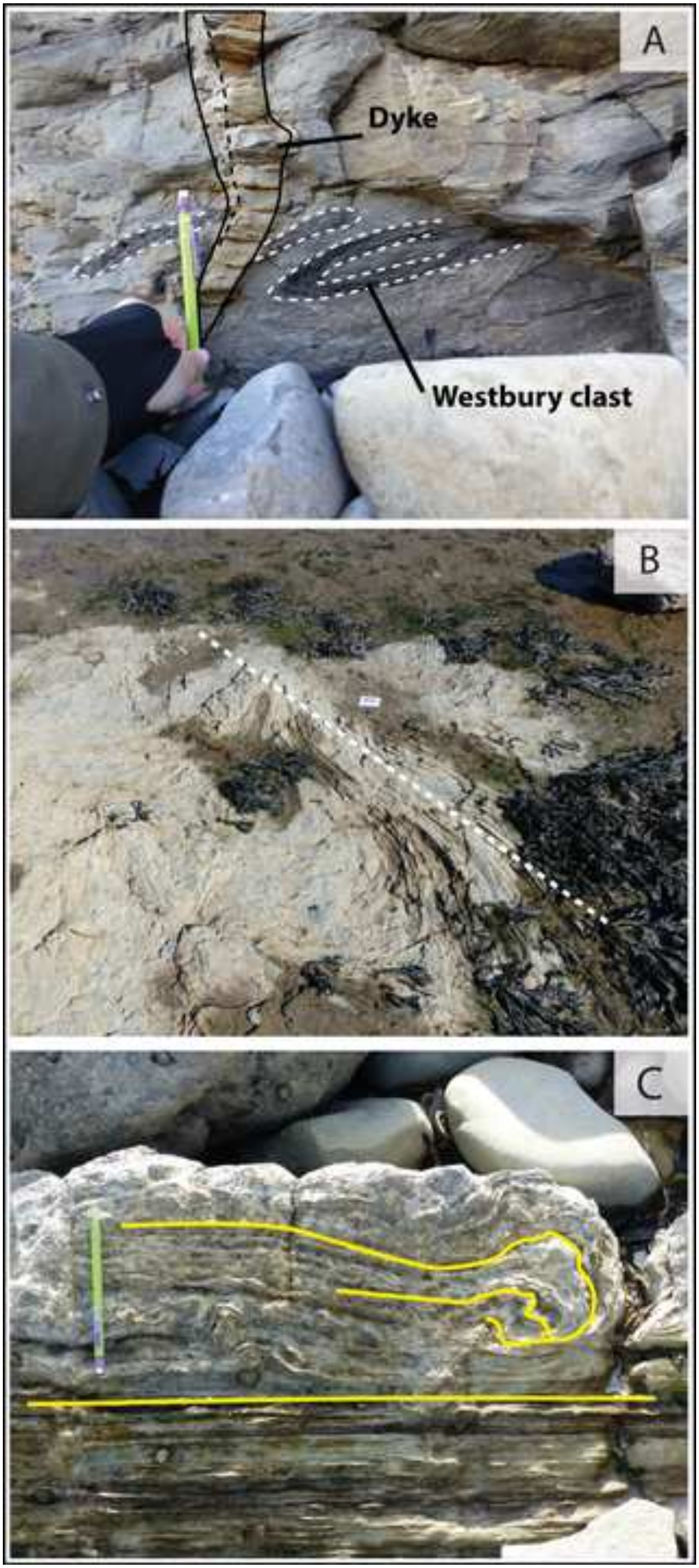


Figure 7

Click here to access/download;Figure;Fig 7.tif

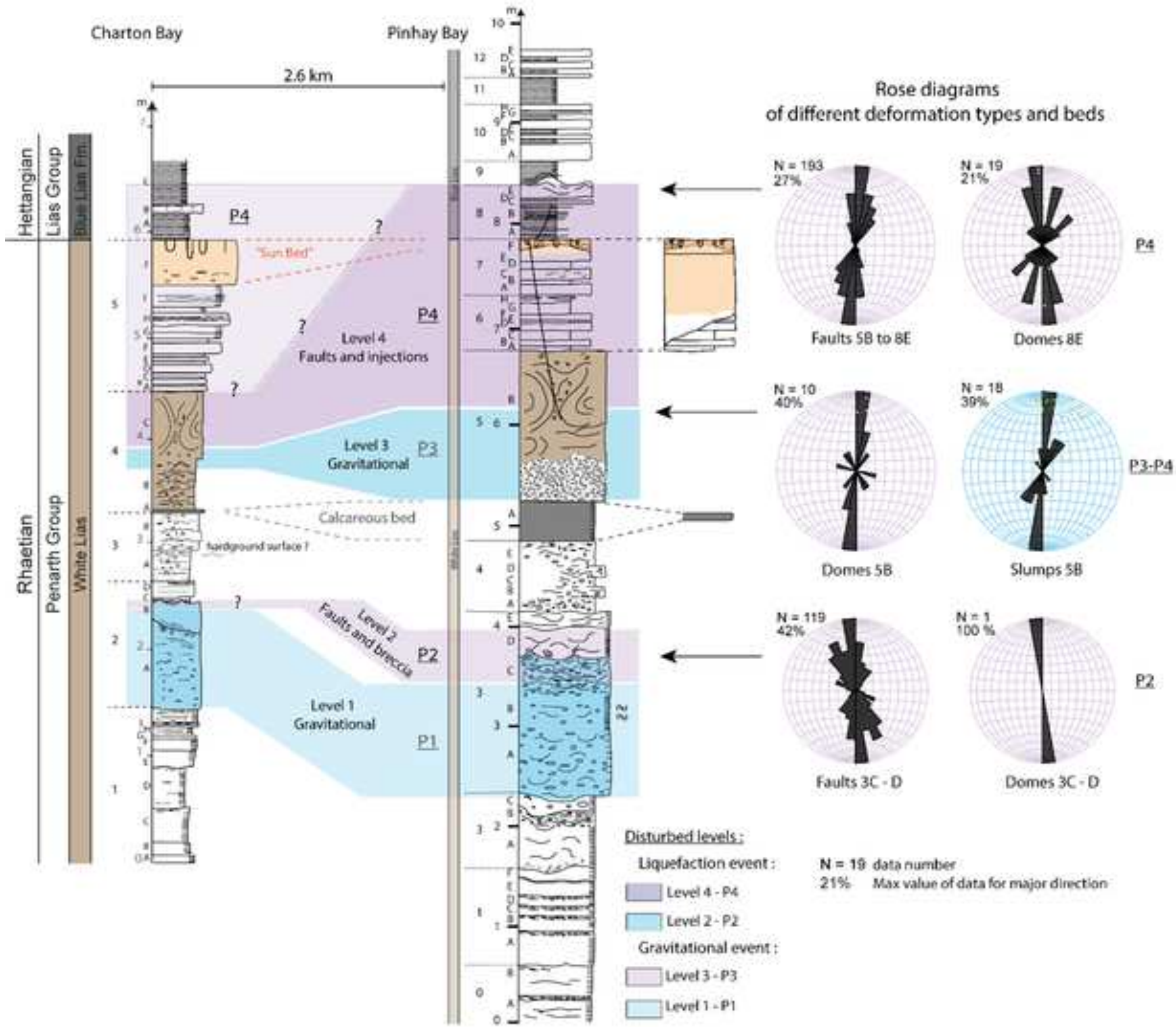


Figure 8

Click here to access/download;Figure;Fig 8.tif

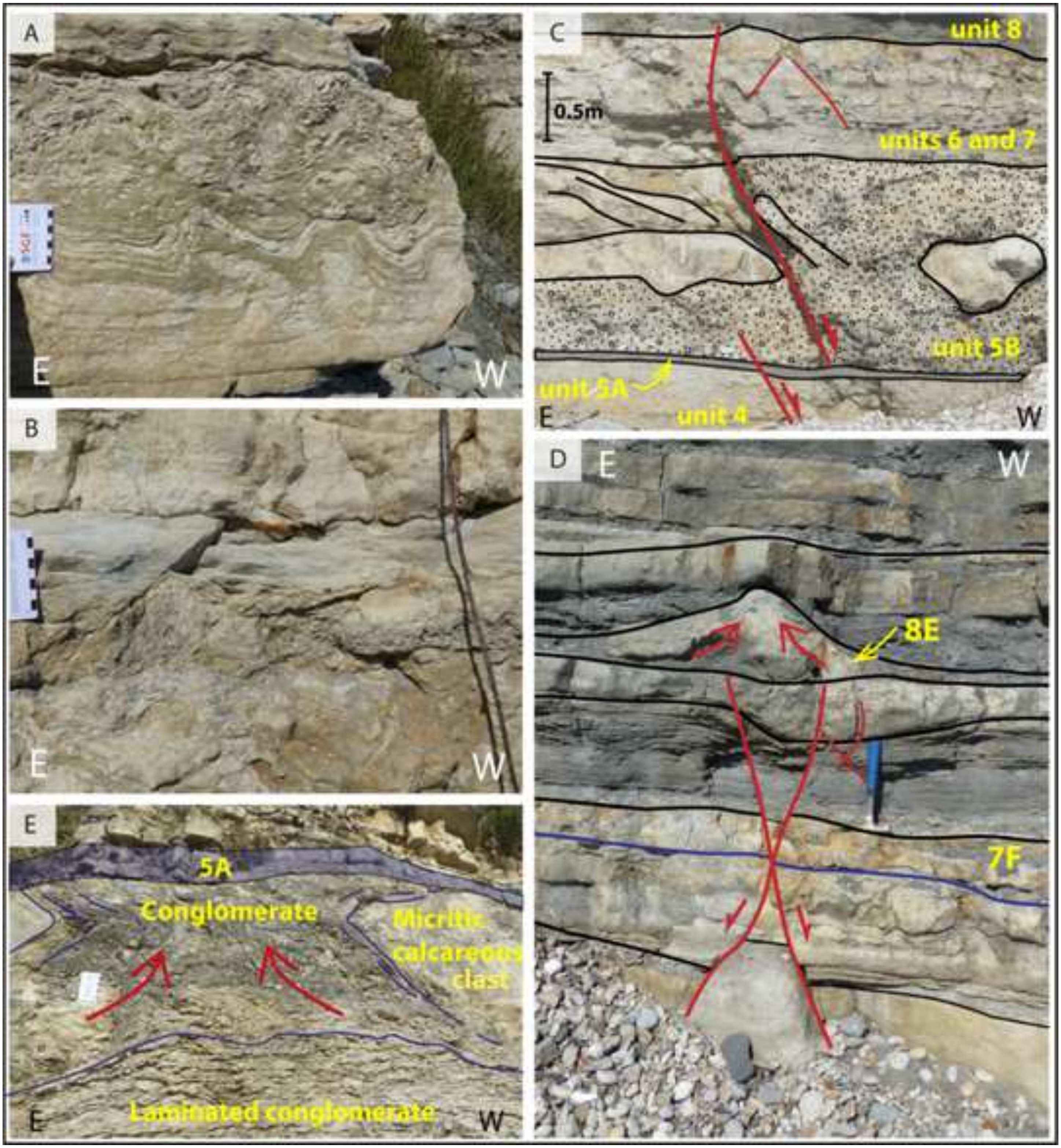
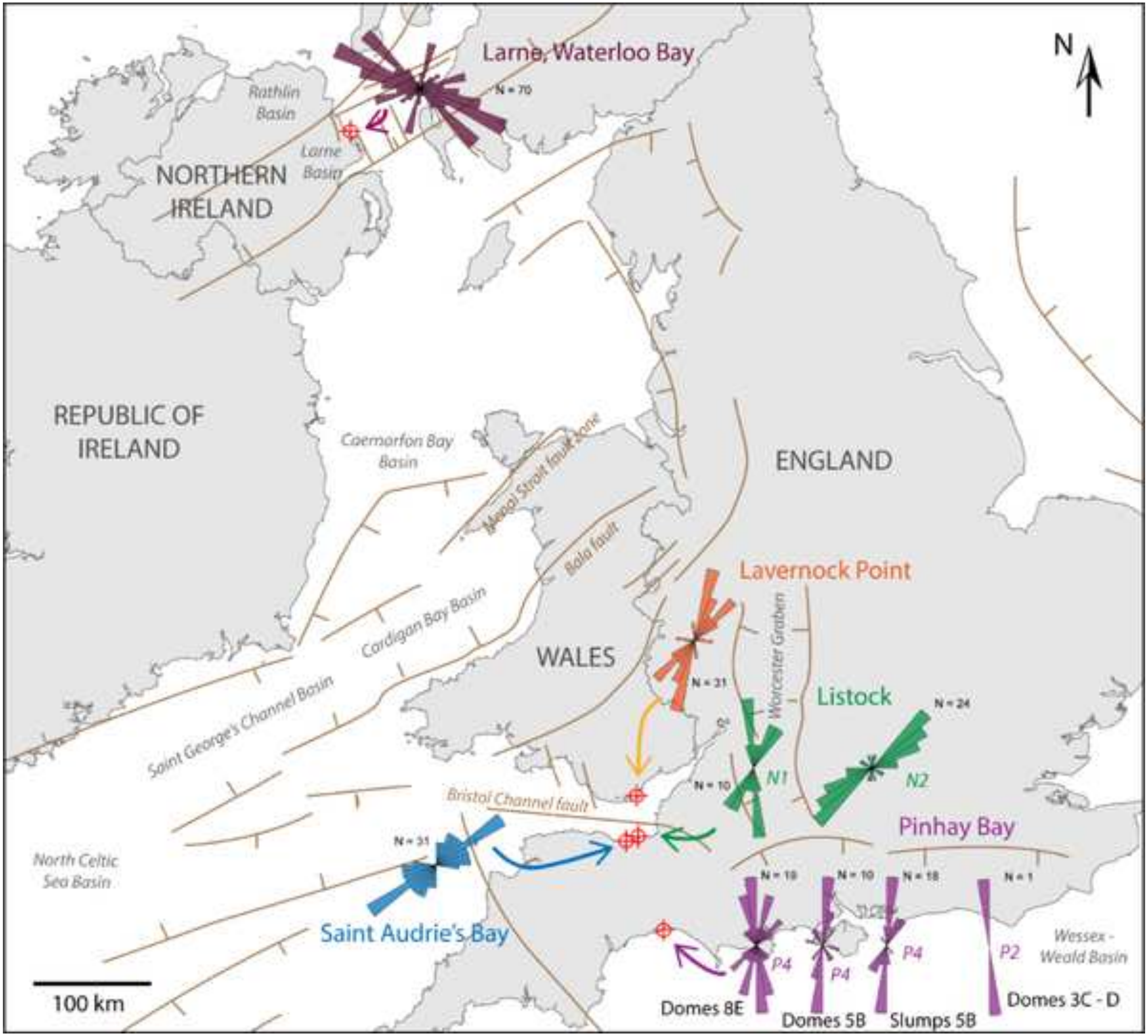


Figure 9



Larne, Waterloo Bay
N = 70
Rathlin Basin
Larne Basin
NORTHERN IRELAND
REPUBLIC OF IRELAND
Caernarfon Bay Basin
Menai Strait fault zone
Bala fault
ENGLAND
WALES
Cardigan Bay Basin
Saint George's Channel Basin
Lavernock Point
N = 31
Worcester Graben
Listock
N = 24
N = 10
N1
N2
Bristol Channel fault
Pinhay Bay
N = 31
North Celtic Sea Basin
Saint Audrie's Bay
N = 10
N = 10
N = 18
N = 1
P4
P4
P4
P2
Wessex - Weald Basin
100 km
Domes 8E
Domes 5B
Slumps 5B
Domes 3C - D
N

Declaration of interests

☒ The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

☐The authors declare the following financial interests/personal relationships which may be considered as potential competing interests: