ROSS AND BUDE FORMATIONS (CARBONIFEROUS, IRELAND AND ENGLAND): REINTERPRETED AS LAKE-SHELF TURBIDITES

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The Ross Formation (Namurian, Ireland) and the near-identical Bude Formation (Westphalian, England), both amply described in the literature, are used by oil companies as deep-sea-fan reservoir analogues. However, the Ross Formation is reinterpreted here, like the Bude Formation in recent publications, to be composed of river-fed turbidites deposited on the wave-influenced northern shelf of a Variscan foreland-basin lake, which also had a southern flysch trough.

Key features of these formations are: (i) two classes of thin (<0.4m) sandstone "event bed" in shale comprising (a) structureless turbidite-like beds, and (b) rippled beds with combined-flow ripples and/or hummocky cross-stratification, neither structure having previously been reported from the Ross Formation; (ii) "trademark" tabular packets (1-10m) of amalgamated event beds which interfinger laterally with mudstones; (iii) sharp packet bases and tops; (iv) rare sinuous channel fills; and (v) rare thick (1-10m) shale units, each containing a thin (cm-dm) fossiliferous band.

The fossil bands are interpreted here as maximum flooding surfaces, reflecting glacioeustatic marine incursions over the lake spill point (sill), forcing the lake to rise and to turn marine or strongly brackish; these bands define Galloway-type depositional sequences 50-100m thick. During eustatic falls, the lake was forced down to sill level, where it perched and turned fresh (desalination). Intervals containing sandstone packets are attributed to the falling-stage and lowstand systems tracts, each packet representing a higher-order lowstand systems tract. Packets are interpreted as tongue shaped, supplied by river-fed underflows. Packet bases (sharp) represent the storm-wave - graded equilibrium shelf profile, glacioeustatically forced to its lowstand position. On this erosion surface were deposited underflow turbidites produced by floods in the catchment. Occasional catastrophic storms on the lake shaved these turbidites and interfingering fair-weather muds back down to the equilibrium level, leaving behind a subsidence-accommodated increment whose surface was sculpted by storm wind and wave currents, forming hummocks, combined-flow ripples and erosional megaflutes. Whenever a river-fed underflow accompanied one of these storms, the resulting highly erosive combined flow carved a sinuous channel on the wave-sculpted equilibrium surface. Sandstone-shale intervals separating the sandstone packets are interpreted as transgressive- and highstand systems tracts. They contain both turbidites and wave-modified turbidites (rippled beds), deposited on the out-of-equilibrium drowned shelf.

A gradual rotation in sole-mark direction with time in both formations is attributed to a reversal of Coriolis deflection as the plate drifted north across the equator, causing underflows (deflected along-shelf geostrophically) to flow first NEwards and then SWwards on an inferred SE-facing shelf.

The lack of evidence for emergence in the Ross and Bude Formations, in spite of the great thicknesses (460m and 1,290m, respectively) of these shallow-water deposits, is attributed to regulation of minimum water depth firstly by the lake sill blocking eustatically-forced exposure, and secondly by storm grading, preventing emergence by sedimentation.

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INTRODUCTION

Discoveries of large oil reserves in deep-sea fan reservoirs surged in the 1990s (Pettingill, 1998). Due to the diversity of fan types (Reading and Richards, 1994) and the high development costs, suitable outcrop analogues are closely studied in order to visualize sand-body geometry and reservoir architecture, and for numerical inputs into reservoir models. One supposed fan analogue is the Ross Formation, of Namurian age (Fig. 1), superbly exposed in cliff sections in County Clare, Ireland (Chapin *et al.*, 1994; Elliott, 2000a; Martinsen *et al.*, 2000). This formation's purported "close analogy with hydrocarbon-bearing successions offshore northern Europe, North America and West Africa has led to research staffs of several oil companies having collected copious field data from the succession" (Martinsen and Collinson, 2002, p. 523). According to Lien *et al.* (2003, p. 113), the Ross Formation is "well suited as an analogue for sand-rich turbidite plays in *passive margin basins*", despite being deposited in a local *"extensional"* basin (emphasis added). However, this assertion is problematic, as the different tectonic setting will inevitably produce differences in catchment size and relief, basin-margin steepness, sediment calibre and depositional environment, among other parameters.

The Ross Formation is virtually identical to the near-coeval (Westphalian) Bude Formation exposed in SW England (Fig. 1; Plates 1, 2; Table 1). The Bude Formation's depositional environment has been debated for several decades (Reading, 1963; Isaac and Thomas, 1998), with two current schools of thought: (i) deep-sea fan (Melvin, 1986, 1987; Burne, 1995, 1998); and (ii) shallow lake floor receiving river-fed turbidites, an interpretation based on the presence of wave-influenced sedimentary structures (Higgs, 1987, 1991, 1994, 1998; *see below*). Wave-influenced structures have never formally been reported from the Ross Formation. However, on two reconnaissance visits, the author has observed numerous examples of symmetrical and near-symmetrical ripples, hummocks and hummocky cross-stratification (HCS), indicating deposition in relatively shallow water. These features are discussed in greater detail below.

In this paper, I propose that the Ross Formation is an unsuitable analogue for marine deep-water fan reservoir rocks because, as will be shown below, it is neither marine (except for some thin fossiliferous bands) nor of deep-water origin. Negative economic consequences of using an inappropriate analogue for oilfield development include, in order of increasing costliness: (i) selection of non-optimum perforation intervals, causing lower production flow rates and lower ultimate recovery; (ii) non-optimum placement, spacing and number of development wells, with the same effects; and (iii) inaccurate predictions of reserves volume and production rates, leading to unwarranted declaration of field economic viability (hence major expenditure such as platforms, development drilling programmes and pipelines) or non-viability.

"CLARE BASIN" TECTONO-SEDIMENTARY EVOLUTION

A foreland-basin setting is well established for the Carboniferous of SW England (Selwood and Thomas, 1988; Gayer and Jones, 1989), but the tectonic setting of the "West Clare Namurian Basin" (Gill, 1979) or "Shannon Basin" (Sevastopulo, 2001) is less well understood.

Foreland basin

According to Ziegler (1990), southward subduction of a Cornwall-Rhenish ocean led to continental collision starting in Late Devonian time. The collision initiated foreland-basin subsidence to the north, including southern Britain and southern Ireland (Ziegler, 1990, encl. 14). In both of these regions, Lower Carboniferous (Dinantian) deposits comprise shallow-marine carbonates ("Carboniferous Limestone"), which pass south into "basinal" shales with cherts and turbidite-like limestone beds (Selwood and Thomas, 1987; Sevastopulo and Wyse Jackson, 2001). This limestone-shale pairing is interpreted here in terms of deposition in an inner (limestone) and outer shelf (shale) setting, adjacent to a flysch trough to the south (cf. Fig. 2), a physiographical configuration typical of an underfilled foreland basin (Allen *et al.*, 1986). The "basinal" deposits grade downward into shallow-marine shales and sandstones (Selwood and Thomas, 1987; Sevastopulo and Wyse Jackson, 2001), consistent with an outer shelf interpretation for the "basinal" strata whose turbidite-like limestones are viewed here as storm beds. Remnants of the supposed flysch trough, disrupted by later thrusting, occur further south, in north-vergent Variscan nappes to the south of Bude (Fig. 2), which contain olistostromes and coarse feldspathic turbidites derived from a north-advancing mountain front (Selwood and Thomas, 1987). The local occurrence of Dinantian volcanics in both the shelf and trough provinces suggests intermittent extension (Ziegler, 1990), possibly related to tectonic escape of the British Isles (Coward, 1993).

The Galway-Mayo and Leinster-Wales-Brabant Highs (Fig. 2) emerged in Dinantian time (Ziegler, 1990; Cope *et al.*, 1999; Sevastopulo and Wyse Jackson, 2001). These highs are interpreted here as the Variscan forebulge, as has been suggested for the Wales portion in Namurian-Westphalian time (Gayer and Jones, 1989). North of the forebulge, Dinantian

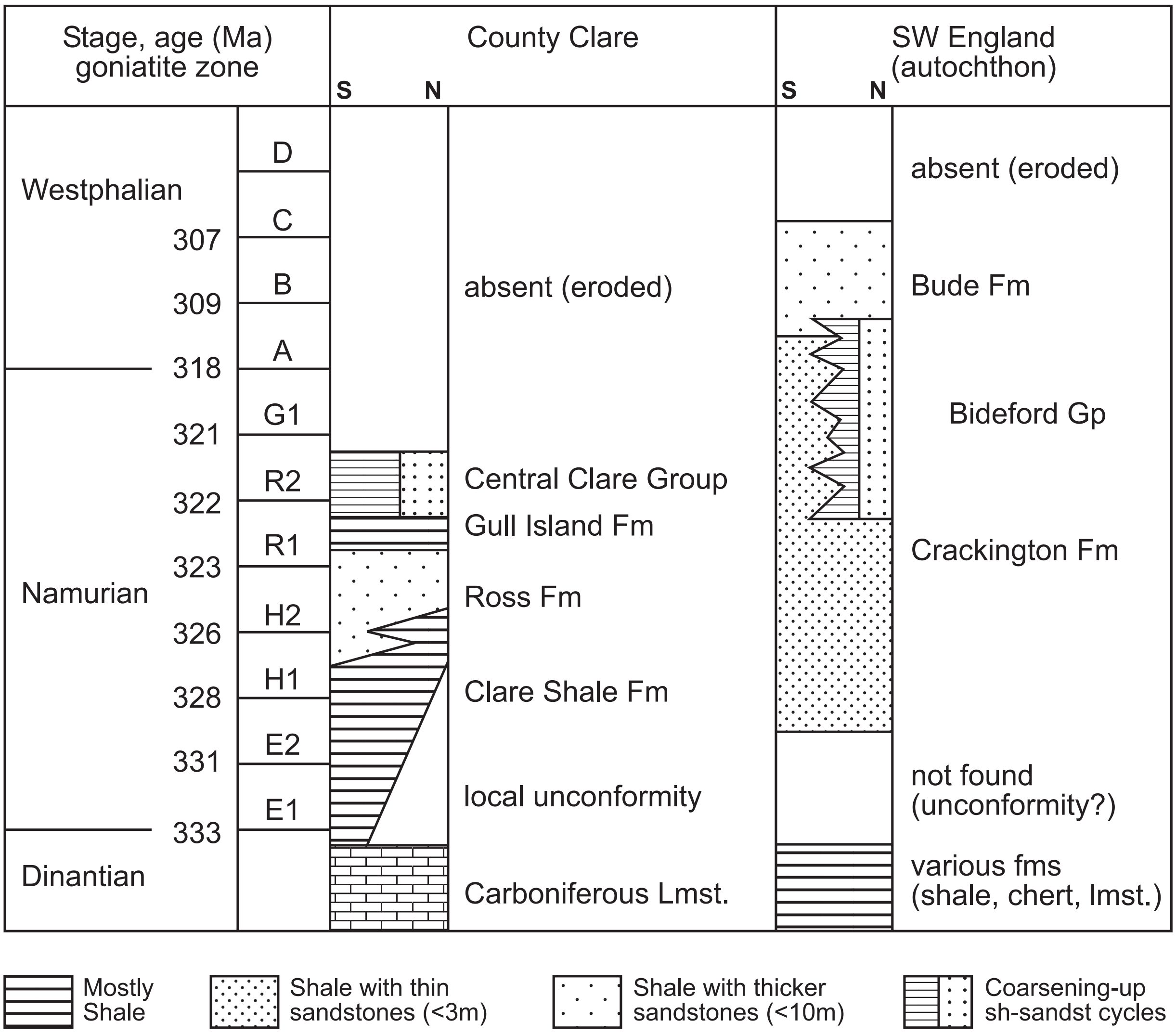

Fig. 1. Carboniferous successions of County Clare and SW England, largely from Ramsbottom *et al.* (1978), Edmonds *et al.* (1979), Freshney *et al.* (1972, 1979), Collinson *et al.* (1991) and Eagar and Xu Li (1993). Absolute ages (Ma) from Harland *et al.* (1990). Vertical scale is time, not thickness. Formation/group thicknesses are: Clare Shale, 10-180m; Ross, 460m; Gull Island, 550m; Crackington, >500m; Bude, 1,290m; Bideford, 820m (Edmonds *et al.*, 1979; Freshney *et al.*, 1979; Collinson *et al.*, 1991; Lien *et al.*, 2003). See text for details of inferred lateral transition of Bideford Group into Crackington and Bude Formations.

extension caused block-and-basin rifting and local volcanism in northern Britain and the northern half of Ireland, forming small carbonate platforms separated by shale basins including the Dublin Basin to the NE of County Clare (Leeder, 1976; Collinson, 1988; Leeder and McMahon, 1988; Strogen *et al.*, 1996).

Previous authors have assumed that County Clare was part of the northern rift province, and was the site of a small Dinantian basin in which deep-water conditions were established in late Dinantian time and persisted into the Namurian until after deposition of the Ross Formation (Collinson *et al.*, 1991; Elliott, 2000a, b; Elliott *et al.*, 2000; Martinsen and Collinson, 2002; Lien *et al.*, 1993). Instead, I propose here that Clare lay within the Dinantian foreland-basin carbonate shelf. Local deepening did occur in the Clare region and was controlled by block faulting (Strogen *et al.*, 1996), consistent with block faulting on forebulge flanks in other foreland basins (DeCelles and Giles, 1996). The controlling faults trend ENE in the Clare region (Strogen *et al.*, 1996), essentially parallel to the forebulge (Fig. 2). Minor volcanism also occurred, but no fault scarps developed (Strogen *et al.*, 1996). By latest Dinantian time, Clare was re-integrated into the regional carbonate platform, with water depths everywhere less than storm wave base (Strogen *et al.*, 1996).

Dinantian-Namurian unconformity

Between the Carboniferous Limestone and the Ross Formation in County Clare is the Clare Shale Formation (Fig. 1). The base of the Clare Shale is an unconformity in north Clare as shown by missing early

Namurian biostratigraphic zones, whereas the time gap is reduced or absent in the south (Hodson, 1954; Hodson and Lewarne, 1961). A thin (<0.1m) phosphatic pebbly bed occurs at the contact in the north (Hodson, 1954). This "condensed bed" (Wignall and Best, 2000, p. 61) is probably a transgressive ravinement lag, as would be expected on an unconformity (Van Wagoner *et al.*, 1990) because (i) the upper surface of the Limestone is "undulatory on a broad scale … interpretable as an erosion surface" (Hodson, 1954, p. 263), with "signs of karstic morphology" (Collinson *et al.*, 1991, p. 227); (ii) the bed contains (phosphatized) limestone pebbles (Wignall and Best, 2000, p. 63), presumably reworked from the Carboniferous Limestone; and (3) the pebbly bed shows *Diplocraterion* and *Rhizocorallium* burrows, suggesting high-energy conditions (Wignall and Best, 2000), consistent with deposition at a transgressing shoreface. The unconformity may reflect block faulting (*i.e.* on the forebulge flank), causing locally reduced subsidence rates, such that subsidence in north Clare was outpaced by a eustatic sea-level fall, resulting in subaerial emergence there. Dramatic north-south thickness variations of Clare Shale and Ross Formation intervals bounded by goniatite markers (Hodson and Lewarne, 1961; Collinson *et al.*, 1991) are consistent with block faulting on ENE-trending faults.

Conversion of the Dinantian marine seaway into a Namurian freshwater lake

Overlying the Carboniferous Limestone in County Clare are the Clare Shale, Ross and Gull Island Formations (Fig. 1). Fossils in these formations are confined to a few thin (cm-dm) bands containing goniatites (Hodson, 1954; Hodson and Lewarne, 1961; Collinson *et al.*, 1991), apart from some cherty beds containing sponge spicules in the Clare Shale (Lewarne, 1963). Following Holdsworth and Collinson (1988), the goniatiferous (and sponge?) bands are interpreted as fully-marine interludes, corresponding to eustatic highstands, in a water body which was fresh or nearly fresh during lowstands. In other words, the water body must have been a sea-level lake (Goldring, 1978), similar to the modern Black Sea and Lake Maracaibo. Similar goniatiferous marine bands occur in the Bude Formation in SW England and also in the underlying Crackington Formation (Fig. 1; Edmonds *et al.*, 1979; Freshney *et al.*, 1979), which resembles the Bude Formation apart from having thinner amalgamated sandstones (Freshney *et al.*, 1979). Freshney and Taylor (1972, p. 468) stated that goniatites in the lower Crackington Formation "tend to be spread through the shales". This could be a false impression given by the closer spacing of goniatite bands in the lower Crackington, where they are only about 10m apart (Butcher and Hodson, 1998), compared to the upper Crackington in which the spacing increases upward from 50m to 150m (Freshney and Taylor, 1972, plate 2). The increase is progressive, possibly reflecting a temporal increase in the rate of subsidence, typical of foreland basins due to encroachment of the thrust load (Allen *et al.*, 1986).

Consistent with the lake interpretation, neither foraminifera nor conodonts have been reported in the Clare Shale or Ross Formations, with the exception of conodonts (age unspecified) recovered from the basal pebble bed of the Clare Shale (Wignall and Best, 2000), which are possibly reworked from the Carboniferous Limestone. Only sparse and very poorly preserved miospore assemblages have been recovered (Fitzgerald *et al.*, 1994; Goodhue, 1996; Goodhue and Clayton, 1999). However, the absence of acritarchs and other marine palynomorphs does not necessarily indicate fresh or brackish water (G. Clayton, *pers. comm.* 2003), because these microfossils are extremely rare in Upper Carboniferous marine rocks in western Europe (Downie, 1984).

Trace fossils (ichnogenus unspecified) are rare in the Ross Formation (Elliott, 2000a, p. 345; Elliott *et al.*, 2000, p. 6), again consistent with the lake interpretation (*see below*). Some "track-marked surfaces" occur in the Cronagort Sandstone of northern Clare (Hodson, 1954, p. 270), which has been assigned to both the Ross Formation (Wignall and Best, 2000) and the Gull Island Formation (Rider, 1974; Collinson *et al.*, 1991; Martinsen and Collinson, 2002). Similarly, trace fossils in the Bude Formation are sparse and of low diversity; only eight ichnogenera have been reported, including crab trackways and fish trails (Higgs, 1988, 1991). The restricted ichnofauna in the Ross and Bude Formations accords with evolutionary models that suggest that few burrowing organisms were adapted to fresh water prior to the Mesozoic (Miller, 1984; Buatois and Mángano, 1995). In contrast, *marine* turbidites of the same age in the USA contain at least 32 ichnogenera (Chamberlain, 1971).

The Clare Shale—Ross—Gull Island and Crackington—Bude successions are interpreted here as having been deposited in the same lake, named Lake Bude by Higgs (1991). By implication, the lake was long-lived (>20Ma; Fig. 1) and large (>300km east-west and north-south, allowing for tectonic shortening; Fig. 2). The conversion of the preceding foreland-basin seaway (in which the Carboniferous Limestone was deposited) into a lake is attributed to the impingement of two Variscan mountain-front salients, somewhere to the west of Clare and the east of Bude, onto the foreland, thus severing the oceanic connection (Higgs, 1991).

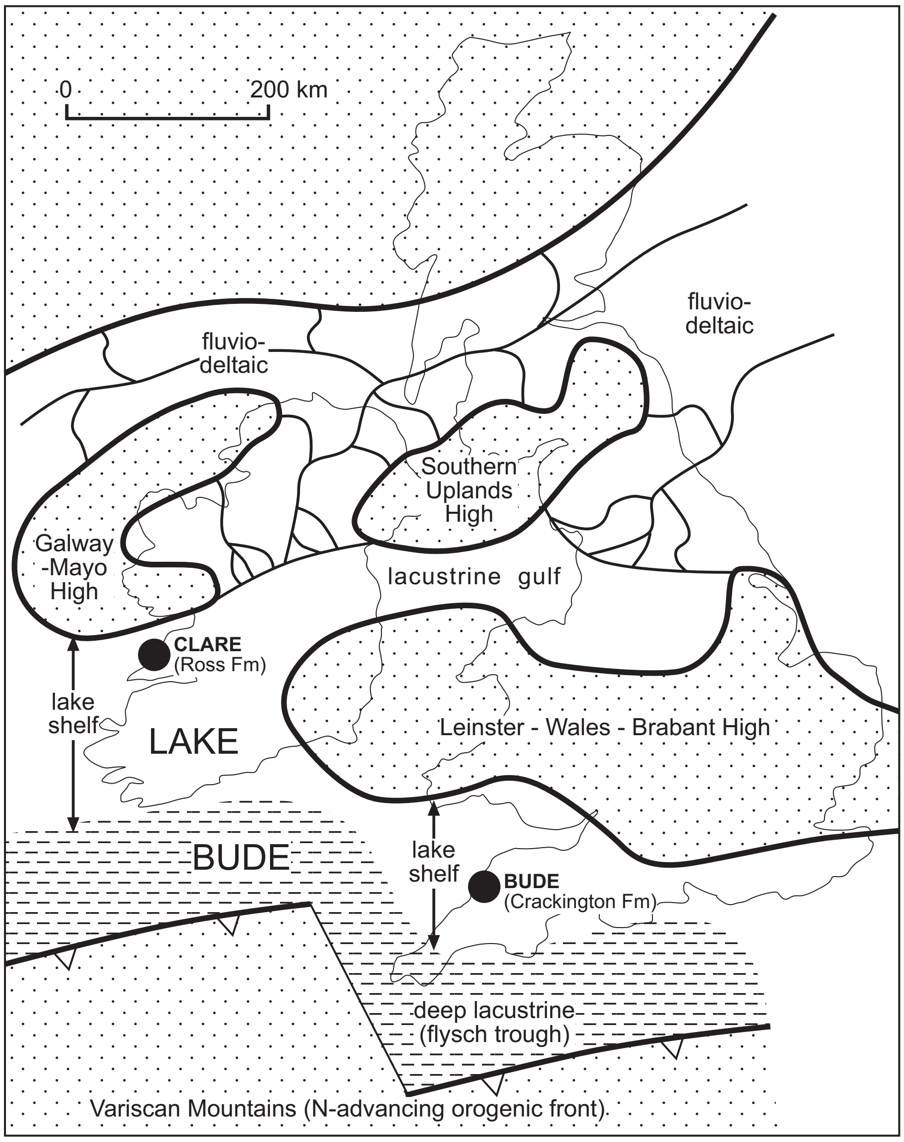

Fig. 2. Mid-Namurian palaeogeography of the British Isles, showing main sedimentary environments and non-depositional highs. A presumed narrow alluvial-coastal fringe surrounding the Leinster-Wales-Brabant High is not shown. Extensively modifed after Ziegler (1990, enclosures 14, 15), Cope *et al.* (1999, map C6) and Sevastopulo (2001, fig. 11.3B). Large dots show approximate positions of County Clare and Bude. Deposition in both of these areas at this time was dominated by river-fed lacustrine turbidites — the Ross and Crackington Formations.

In common with all lakes, Lake Bude must have had a sill to impound the water (Talbot and Allen, 1986). The sill was probably located at one of the mountain-front salients. Glacioeustatic sea level fluctuations, characteristic of the Late Carboniferous (Frakes, 1979), were capable of drowning the sill (Higgs, 1991, fig. 20). During eustatic lowstands, sea level fell below the sill, which accordingly became a spill-point outflow (incised gorge). At such times, the lake remained brimfull due to the equatorial humid climate with inflow exceeding evaporation, and would have undergone "desalination", with water becoming fresh or almost fresh (Holdsworth and Collinson, 1988, p. 137). When eustatic sea level rose again, the sill would eventually be overtopped by the sea, forming a strait similar to those connecting the Black Sea and Lake Maracaibo to the ocean. Further deepening of the strait (and the lake) forced by the rising sea level would allow a sea-water wedge to enter the lake, which accordingly would turn increasingly brackish, becoming a marine gulf during extreme eustatic highstands.

Palaeobathymetry of Lake Bude

Within the context of Lake Bude, the Clare Shale is interpreted here as having been deposited on a drowned (sand-starved) shelf, in relatively shallow water (50-200m), as opposed to the "several 100 m?" proposed by Wignall and Best (2000, p. 76). Evidence for shelf deposition is that (i) the Clare Shale overlies

Dinantian shelf carbonates; (ii) it locally has a basal transgressive lag; and (iii) it grades laterally and upward into the Ross Formation, which contains wave-influenced sedimentary structures (*see below*). Consistent with a south-facing shelf, the flysch trough persisted to the south as shown by the presence of coarse, feldspathic turbidites to the south of Bude which reach up into the early Namurian (Selwood and Thomas, 1986). The shelf-trough configuration is inferred to have persisted through Namurian and Westphalian A-B time (Figs. 1, 2), based firstly on the presence of wave-influenced structures in the Ross—Gull Island and Crackington—Bude successions (*see below*); and secondly on the fine and very fine grain-size, quartzose mineralogy and lack of north-flowing palaeocurrents in the Crackington and Bude Formations (Edmonds *et al.*, 1979; Freshney *et al.*, 1979; Melvin, 1986), despite the north-advancing mountain front to the south (Selwood and Thomas, 1988), consistent with the model that these formations were deposited on a shelf while orogen-derived detrital material was trapped in an intervening trough (Higgs, 1991).

PRIMARY SEDIMENTOLOGICAL FEATURES

Any model for the Ross and Bude Formations must account for the presence of the following key sedimentological features (references in Table 1):
(i) two classes of sandstone "event bed" in shale, firstly turbidite-like structureless beds and secondly rippled and hummocky beds; (ii) characteristic tabular packets (1-10m) of amalgamated event beds; (iii) sharp packet bases and tops; (iv) rare, metre-scale, thickening-upward successions; (v) rare small channels; (vi) a rare flaggy siltstone facies with fish trails and crab trackways; (vii) rare slump beds; and (viii) rare thick (1-10m) shale units, each containing a thin (cm-dm) fossiliferous band, in some cases demonstrably marine (goniatites). These features are discussed briefly in turn below.

Structureless beds

Description

Thin (< 0.4m), dominantly structureless sandstone beds are abundant in the Ross and Bude Formations (Table 1). These beds are composed of fine- or very fine-grained sandstone, typically grading into siltstone or very fine sandstone in their upper 0.01-0.02m which may show parallel lamination. Sole marks include flutes, grooves and longitudinal scours (Higgs, 1991). The across-current extent of these sandstone beds is unknown. The downcurrent extent could be as much as 20-30km, given that some parts of the Bude Formation outcrop are at least 20km from the nearest possible lake shore, after correcting for tectonic shortening (Higgs, 1991); and the Loop Head outcrop of the Ross Formation is 10km long parallel to the northeastward palaeoflow, while equivalent strata 50km to the NE are shales (Clare Shale at Lisdoonvarna; Wignall and Best, 2000, figs 1, 3).

Interpretation

Following Higgs (1991, 1998), these beds are interpreted as lake-shelf turbidites deposited by slow (< 1m/sec), river-fed underflows generated by exceptional rainfall in a distant area of the catchment, without storm waves on the lake. Each underflow lasted perhaps a few days. An underflow of 1 m/sec would be incompetent to transport sediment coarser than medium sand in suspension (Sundborg, 1967, fig. 1), consistent with the fine grain size of Ross and Bude sandstones (Table 1). In contrast, coeval deep-water turbidites in the extensional basins of northern Britain are coarser grained (*e.g.* the Shale Grit; Collinson, 1988). These deposits are attributable to high-velocity turbidity currents accelerated down a high-gradient basin slope and generated by river inflow or by slumping. A 1 m/sec underflow is also consistent with flutes (Table 1; Higgs, 1991).

The lack of parallel lamination throughout much or all of a bed could suggest rapid fall-out from suspension, inhibiting traction (Lowe, 1982). This would imply that the underflow initially travelled a long distance before deposition began, to account for the combination of thinness (cm-dm) and probable great lateral extent (J. Southard, *pers. comm.,* 2003). The failure of the flows to produce current-ripple lamination as they decelerated may also reflect rapid fall-out from suspension, preventing ripple nucleation (Lowe 1988, fig. 3, upper tier).

Rippled and hummocky beds

Description

Also occurring in the Ross and Bude Formations are thin (<0.4m) beds of fine- or very-fine sandstone capped either by hummocks, or by symmetrical or near-symmetrical ripples of variable size (large or small) and regularity (2D or 3D) (Plates 3-8). Near-symmetrical small and large ripples also occur in the Gull Island Formation (Plates 9, 10) overlying the Ross Formation. Large ripples do not exceed 1m spacing and 0.15m amplitude. There are no larger, steeper, strongly asymmetrical "dunes" or "megaripples" (Ashley, 1990), the grain size being too fine (Myrow and Southard, 1991).

These structures have not previously been reported from the Ross or Gull Island Formations. In the Bude Formation, near-symmetrical small ripples and symmetrical large ripples were reported by Higgs (1991).

	Bude Fm	Ross Fm
Three main facies: 1. shales 1-10m 2. interbedded cm-dm sandstones and shales 3. tabular amalgamated sandstones 1-10m	2,5,7	1,4,6
Sandstone beds are fine- and very-fine grained	2,5,7	4,6
Amalgamated sandstone component beds are mostly thin (cm-dm)	5	1,4
Amalgamated sandstones can de-amalgamate laterally; each component sandstone bed occupies a flat-based channel <0.4m deep, carved in mud	5	6,8,9,12
Poor vertical bed-thickness successions. Instead, sandstone 'packets' alternate abruptly with interbedded sandstones and shales	2,5,7	4,6
Most sandstone beds are massive, or massive with minor (mm-cm) parallel-laminated top	2,5,7	4,6
Rare sandstone-filled channels 1-10 m deep, containing turbidites and/or cross-bedded sandstone (dm sets) and/or lateral accretion and/or mudclast conglomerate	7	4,6,8,10,11,12
Sandstone beds with flutes	2,5,7	1,4,6
Mudstone-draped megaflutes	nr	6,8,9,12
Undulating mudstone-draped scours	5	4,8,9,10
Beds capped by near-symmetrical ripples	3,5	nr
Beds with hummocky cross stratification	5	nr
Marine fossils confined to rare bands in Facies 1	2,5,7	1,4,6
Trace fossils scarce	3,5,7	8,9
Slump beds	2,5,7	1,4,6
Slurry beds	5,7	6

Table 1. Facies and sedimentary structures shared by the Ross Formation and Bude Formation. Key to references: 1, Rider 1974; 2, Melvin 1986; 3, Melvin 1987; 4, Collinson *et al.* 1991; 5, Higgs 1991; 6, Chapin *et al.* 1994; 7, Burne 1995; 8, Elliott 2000a, b; 9, Elliott *et al.* 2000; 10, Martinsen *et al.* 2000; 11, Wignall and Best 2000; 12, Lien *et al.* 2003. nr = not reported.

The internal structure of these beds is correspondingly variable, including HCS (symmetrical or asymmetrical; Plates 4, 8), large-ripple trough cross-stratification (dm scale; climbing and non-climbing), and small-ripple cross-lamination (climbing and non-climbing). Any of these types of lamination can be faint and difficult to see (Plate 8). A basal division with blurred horizontal or low-angle lamination may be present and can dominate a bed.

HCS has previously been reported in the Bude Formation (Higgs, 1991), but never in the Ross Formation. Lamination "comparable to" HCS was reported in the Bude Formation by Burne (1995, p. 114). The Crackington Formation also reportedly contains HCS (Eagar and Xu Li, 1993, fig. 10).

Interpretation

Again following Higgs (1991, 1998), these beds are interpreted as the deposits of simultaneous river-fed underflows and storm waves, *i.e.* combined flows. The

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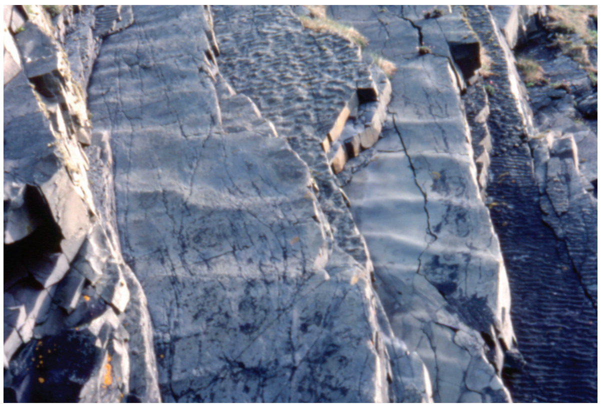
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Plate 1. Ross Formation at Loop Head, Co. Clare, Ireland. Tabular sandstone packets are prominent on all visible cliff faces. Note chevron fold in foreground.

Plate 2. Bude Formation at Upton, near Bude, Cornwall (England). Stratigraphic top to right. Tabular sandstone packets are prominent. A central cluster of four thick (up to 7m) packets is surrounded by thinner-packeted intervals. Cliff height about 45m. Backpack and notebook for scale near cliff base, resting on second thick packet. The third packet appears to de-amalgamate up the cliff, and becomes thinner, possibly due to differential compaction. See measured section by Higgs (1991, fig. 13).

Plate 3. Bude Formation, Greenway Beach, near Morwenstow. Rippled tops of sandstone beds encased in shales. Two bedding planes show large symmetrical ripples with bifurcations and slightly sinuous crest-lines; ripple wavelength is approximately 50cm. A lower bedding plane (bottom right) shows small, near-symmetrical, straight-crested ripples. Note constant crest-line orientation of large and small ripples.

Plate 4. Bude Formation, Upton. Sandstone bed with hummocky top and hummocky cross-stratification (HCS), overlying grey silty mudstone. Coin diameter 3cm.

Plate 5. Ross Formation at Rinevella. Interbedded sandstones and shales. Two sandstone beds at centre have hummocky tops. Sandstone unit above is probably amalgamated.

Plate 6. Ross Formation, Rinevella. Sandstone bed below hammer handle has hummocky top. Thick sandstone unit touched by hammer-head shows undulatory partings, which are possibly set boundaries of blurred HCS (see text). Note lack of obvious bed-thickness successions.

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Plate 7. Ross Formation, Rinevella. Two sandstone beds at centre have hummocky upper surfaces. Note 15cm scale on lower bed above loose boulder (close-up next photo).

Plate 8. Close-up of the lower hummocky bed in Plate 7. Faint low-angle lamination (e.g. right of scale) is interpreted here as HCS. Upper surface (receding in this view) shows weathered ripples which are slightly sinuous and nearly symmetrical.

Plate 9. Gull Island Formation, near Gull Island. Steeply dipping upper surface of a sandstone bed. Person for scale at extreme upper left. A megaflute occurs at right centre "pointing" downward (palaeoflow upward). Smaller megaflutes are visible higher up. On the left, the same surface shows large, sinuous-crested ripples which are only slighly asymmetrical. See close-up photo by Elliott (2000a, fig. 6).

Plate 10. Gull Island Formation, near Gull Island. Interbedded sandstones and shales dipping toward viewer. Several sandstone bed tops to left and right show slightly sinuous, near-symmetrical ripples.

Plate 11. Bude Formation, Northcott Mouth. Amalgamated sandstone resting abruptly on shale. Note flutes (or flaring gutters?) at base of sandstone. Iron-oxide encrusted joint surfaces obscure lamination (if any) in lower half of sandstone. Upper half shows vague low-angle lamination which is possibly blurred HCS.

Plate 12. Bude Formation, Maer. Stratigraphic top to right. Sea stack shows a channel containing dm-scale cross-stratified sandstone (note trough shapes on weathered surface behind person; cf. Burne, 1995, figs. 34B, 36). Channel wall obliquely truncates the tabular sandstone unit slightly left of centre and flattens into the (bedding-parallel) channel floor about 1m left of person's feet. Lower part of channel fill is thin-bedded sandstone and shale.

diversity of internal structures and bedforms reflects the complexity of deposition from combined flows (Myrow and Southard, 1991). Beds of this kind have been termed "wave-modified turbidites" (Myrow *et al.*, 2002). All previous publications on the Ross Formation have inferred the ripples to be asymmetrical "current ripples". The weakly laminated basal interval could reflect wave-induced liquefaction and water escape, blurring the lamination (Owen, 1987).

Sandstone packets

Description

Structureless sandstone beds are commonly amalgamated to form tabular sandstone packets up to 10m thick, characteristic of both the Ross and Bude Formations (Facies 3 in Table 1). HCS is seldom clearly expressed inside packets, but many packets have one or more thin (dm) intervals with faint lamination (Plate 11), commonly oversteepened, which is possibly HCS that has been deformed and/ or partially blurred by sand-on-sand loading and liquefaction, possibly induced by earthquakes (*see below*).

Packets can de-amalgamate laterally (Higgs, 1991, p. 453 and figs 8, 9; Chapin *et al.*, 1994, p. 58; Elliott, 2000a, p. 352; Lien *et al.*, 1993, p. 126), due to the presence of "stepped scours" and "bedding-plane scours" with low (cm-dm), sub-vertical walls (Chapin *et al.*, 1994, fig. 7). These features suggest that each sand bed within a packet occupied (and overspilled) a flat-based, mud-walled, shallow (cm-dm) scour. Within any packet, successive scour walls can occur within a zone less than 50m wide (Higgs, 1991; Lien *et al.*, 2003, fig. 13), suggesting that the constituent beds in a packet are approximately superimposed. Packets are thought to be a few kilometres wide transverse to the palaeoflow direction (Higgs, 1991). The flow-parallel length could be as much as 20-30km, similar to that of the component beds. Some intervals are markedly sandier than adjacent intervals above or below, reflecting a greater average packet thickness.

Such intervals are tens to hundreds of metres thick (Plate 2; Higgs, 1991, fig. 10). For example, the upper Ross Formation (290m thick) is sandier than the lower (170m; Lien *et al.*, 2003, fig. 4).

Interpretation

Packets are thought to reflect lake lowstands (*see integrated interpretation below*). Each packet is interpreted as tongue-shaped (Higgs, 1991), emanating from a former river mouth. Tongues are inferred to bend shore-parallel, deflected by the Coriolis effect (*see below).* Tongues which were fed by different rivers may have merged along-shelf. Possible reasons for clustering of thicker packets are addressed below.

Mud-draped scours and megaflutes

Description

In the Ross Formation, Elliott (2000a, b) reported laterally extensive (> 1km) mud-draped erosion surfaces on sandstone which undulate on a medium scale (m-10s m) with low relief (dm-m). Similar low-relief, undulatory, mud-draped scours occur in the Bude Formation (Higgs, 1991, pp. 452-453 and fig. 7). These erosion surfaces are ornamented with megaflutes in the Ross Formation (Elliott 2000a, b). Megaflutes were also recognized by Collinson *et al.* (1991) who referred to them as spoon-shaped mud-filled scours, and by Chapin *et al.* (1994). Dimensions are 1-45m across and 0.5-3m deep (Collinson *et al.*, 1991; Elliott, 2000b), similar to those of mud-draped "concave-up scours" in the Bude Formation (Higgs, 1991, p. 453), which may be megaflutes exposed in cross section. The vertical spacing of megafluted erosion surfaces in the Ross Formation varies from 5m to several tens of metres (Elliott, 2000b). Megaflutes also occur in the Gull Island Formation (Plate 9).

Interpretation

Megaflutes were attributed by Chapin *et al.* (1994) to bypassing of sediment in an erosive deep-water turbidity current. The same interpretation was later applied to the undulating megafluted surfaces (Elliott, 2000a, b; Martinsen *et al.*, 2000). In contrast, these surfaces are attributed here to shelf grading by catastrophic storms (*see below*).

Channels

Description

The Ross Formation contains narrow (100-300m), non-leveed, sandstone-dominated channel fills up to 10m thick (Collinson *et al.*, 1991; Chapin *et al.*, 1994; Elliott, 2000a, b; Elliott *et al.*, 2000; Martinsen *et al.*, 2000; Wignall and Best, 2000, 2002; Lien *et al.*, 2003). Channels make up about 10% of the entire Ross Formation (Collinson *et al.*, 1991) and are confined to the upper, sandier part (Lien *et al.*, 2003). The channels cut down from the top or base of a (tabular) sandstone packet, or from an internal bedding surface (Elliott, 2000a, b; Lien *et al.*, 2003), and can reach down into the packet below, thereby linking two packets. The portion of a packet which overlies an internal or basal-surface channel has been called a "channel wing" (Elliott, 2000a, b; Elliott *et al.*, 2000), perhaps misleadingly, as it post-dates the channel. Channel sides can be either curved or stepped. Five kinds of channel fill can occur singly or in various combinations: four of them comprise thin (cm-dm), structureless sandstone beds or interbedded thin sandstones and mudstones, which can be either vertically- or laterally accreted in both cases; and the fifth is composed of sandstone beds with decimetre-scale trough cross-stratification. Mud clasts, lithologically resembling mudstone beds exposed in channel walls, can occur at the channel base and within laterally-accreted sandstone beds. Lateral accretion indicates that channels are sinuous, with a curved "point bar" side and a stepped "cut bank" side (Lien *et al.*, 2003, fig. 24). Similar channels occur rarely in the Bude Formation (Burne, 1995, figs 34A, 36). Like the Ross examples, the Bude channels have stepped or curved sides, basal mud clasts and variable fill, including cross-stratified sandstone (Plate 12) and structureless sandstone beds accreted vertically and laterally (Burne, 1995, pp. 117, 119, figs 33-36).

Interpretation
The channels were interpreted by Elliott (2000a, b) as having been excavated by the same turbidity currents as were envisaged to have carved the mud-draped scours; they were subsequently filled by later, weaker turbidity currents. Lien *et al.* (2003) also attributed the channel fills to a series of turbidity currents, but did not discuss how the channels were cut. Channel cutting is interpreted here (*see below*) to have occurred during lowstands, by river-fed underflows reinforced by storm wind- and wave currents.

INTEGRATED DEPOSITIONAL AND SEQUENCE STRATIGRAPHIC MODEL

No sequence stratigraphic interpretation of the Ross or Bude Formations has been published previously. These formations are interpreted here as stacked lake-shelf sequences whose deposition was controlled by glacioeustatic fluctuations. The most convenient sequence boundaries by which the succession can be divided are maximum flooding surfaces (MFS) (Galloway, 1989). Exxon-type sequence boundaries (Van Wagoner *et al.*, 1988) are difficult or impossible to recognize in the Ross and Bude Formations because lowstand subaerial unconformities are absent, the lake sill having kept the lake brimfull at lowstands, thereby preventing subaerial exposure of the Ross and Bude shelf areas *(see below*). This contrasts with *marine* shelves, where forced regressions can pass the shelf edge. Despite this peculiarity, the Ross and Bude sequences can be described using the conventional nomenclature of highstand-, transgressive- and lowstand systems tracts (HST, TST, LST), together with the recently-defined falling stage systems tract (FSST; Plint and Nummedal, 2000).

The model

The fossiliferous bands are interpreted here as MFS's, reflecting glacioeustatic marine incursions over a lake sill, forcing the lake to rise and to turn marine or strongly brackish; these define Galloway-type depositional sequences. Six goniatiferous (marine) shale bands are known in the Ross Formation, including those at the base and top (Collinson *et al.*, 1991, fig. 8), compared to nine in precisely the same biostratigraphic interval in northern England (Holdsworth and Collinson, 1988), suggesting that some bands in Clare await discovery and/or are represented by shale bands that are not fully marine. The time interval between successive Namurian goniatiferous bands is about 185,000 years on average (Holdsworth and Collinson, 1988). Therefore, the duration of the Ross Formation is approximately 1.5Ma (eight 185,000-year sequences bounded by the nine shale bands of northern England; cf. Fig. 1), and the rate of subsidence is about 300m/Ma (the total thickness is 460m according to Lien *et al.*, 2003). In the Bude Formation, goniatiferous shale bands mark the base and top (Freshney *et al.*, 1979, fig. 3). The time significance of the Bude bands is unclear, as goniatiferous bands are less frequent in the Westphalian than the Namurian, and are less regularly spaced (Ramsbottom *et al.*, 1978). Other bands in the Bude Formation contain fish but no goniatites (Freshney *et al.*, 1979), and are interpreted as brackish (Higgs, 1991). Six fossiliferous bands have been found in total, spaced 50-600m apart (Freshney *et al.*, 1979, fig. 3). The irregular spacing may reflect structural repetition or omission, or may indicate that more fossil bands await discovery, possibly because they are dispersed in slump beds.

Sandstone packets are confined to a central interval estimated to occupy 80-90% of the total thickness of each sequence (*e.g.* Higgs, 1991, fig. 10; Lien *et al.*, 2003, fig. 4). This interval is interpreted as the combined FSST-LST. Individual packets are interpreted to reflect higher-order lake lowstands, caused by higher-order eustatic cycles (possibly compounded Milankovitch 40,000 and 20,000-year cycles; cf. Berger, 1988), which left the lake perched at sill level whenever sea level fell below the sill. During these lowstands, the lake became fresh, favouring river-fed underflows induced by floods in the catchment (monsoons or "super-monsoons"?). Fresh (lake) water is much more prone to underflow occurrence than sea water, which requires underflow sediment concentrations at least 50 times greater (Mulder and Syvitski, 1995). Thus in lakes, more river-borne sand can be delivered directly offshore, as opposed to being dumped at the river mouth and redistributed by longshore drift. Consequently, *aggradation* of the lake floor is augmented at the expense of *progradation* of the lake margin.

The contact between the FSST and the LST is difficult to locate (cf. Plint and Nummedal, 2000). A cluster of thicker-than-average packets may be recognizable (Plate 2) and is assigned here to the LST. The thickening is interpreted to reflect the increasing duration of successive higher-order lake lowstands, as more of each successive higher-order eustatic cycle lay below the sill during the falling limb of the lower-order cycle. In fact, one or more higher-order cycles may have been entirely below sill level during the (lower-order) lowstand, potentially producing composite, overthickened packets.

The sharp base of each sandstone packet (Plate 11) is interpreted to be wave cut, representing the storm-graded shelf equilibrium profile (Seilacher, 1982), which was forced down to its lowstand position by the falling sea/lake level (cf. "regressive surface of marine erosion"; Plint and Nummedal, 2000). Erosional gutters containing sandstone with HCS, common below wave-cut regression surfaces (Plint and Nummedal, 2000), are rare in the Bude Formation (Plate 11; Higgs, 1991), and are unreported in the Ross Formation, possibly because the weak ionic cohesion of fresh-water muds (Van Olphen, 1977) inhibited gutter development by promoting the resuspension of fluid mud.

During the lower-order lowstands, empty incised valleys bordered the lake, each containing an axial river flowing into the lake. The valleys were filled during the ensuing transgression, forming laterally extensive (km-10s km) fluvial sandstone bodies like those in the Central Clare Group (Fig. 1) and the Bideford Group near Bude (*see below*). These sandstone bodies have been interpreted previously as: distributary channels (Elliott, 1976); "widespread fluvial deposits controlled by sea-level changes" (Xu Li, 1990); and incised-valley fills (Elliott and Pulham, 1991; Elliott *et al.*, 2000).

In front of each river, a tongue-shaped (km x 10s km?) sand packet accumulated on the shelf equilibrium surface by the deposition of successive flood underflow beds. Mud layers deposited between floods, preserved on tongue flanks, were eroded in the tongue itself leading to sand-bed amalgamation. Less frequent shelf-grading events, attributed to catastrophic storms (centennial or millennial?), shaved these sands and muds back down to the equilibrium surface, leaving an increment accommodated by subsidence (Higgs, 1991) whose top was reworked by storm wave and wind currents, producing megaflutes, hummocks and large and small ripples (cf. Plate 9) possibly underlain by a thin (dm) interval with HCS.

In view of the near-equatorial setting (Scotese and McKerrow,1990), storms were probably not hurricanes (Marsaglia and Klein, 1983). Storms of other origin do occur at the equator, although much less frequently than at higher latitudes (Petterssen, 1958).

Shelf-grading events generally did not produce winnowed lags because lag-forming materials are scarce in the Ross and Bude Formations (*e.g.* no shells in fresh-water intervals; no palaeosoil nodules because the shelf was never exposed (*see below)*; no pebbles because the area was beyond the reach of pebbly flows). However, the author has seen one example, in the Bude Formation, of gutters containing nodules (fossiliferous ?) , interpreted to have been winnowed from one of the fossiliferous shale bands.

As well as the clustering of thicker packets in the LST, as described above, average packet thickness can change markedly from one sequence to the next. Two interrelated interpretations derived from this observation are that:

(i) the packets (tongues) are shelf-parallel, due to Coriolis deflection of the underflows (*see below*); and

(ii) the thicker-packeted sequences are closer to the sand-supplying river mouth, in an along-shelf sense.

The change in average packet thickness from one sequence to the next is thus interpreted to reflect a shift in river (*i.e.* incised valley) position.

River-fed underflows accompanying the storm wind and wave currents resulted in highly erosive combined flows carving channels on the wave-worked equilibrium surface. Channels are confined to sequences with relatively thick packets, indicating that the channels were confined to the proximal reaches of sand tongues. Channel sinuosity may reflect non-alignment of the unidirectional and the oscillatory flow components of the combined flow. The channel would have confined subsequent underflow beds until it was filled. The trough cross-stratification within the channel (see Plate 12) is interpreted here as a combined-flow variety (Myrow and Southard, 1991).

Sandstone-shale intervals separating the sandstone packets are interpreted as higher-order transgressive and highstand systems tracts. These intervals contain both turbidites and wave-modified turbidites (rippled and hummocky beds) deposited on the shelf, which was now out of equilibrium (drowned) until the next lowstand. Uncommon, metre-scale thickening-up intervals reported in the Ross and Bude Formations (Burne, 1995; Lien *et al.*, 2003) are interpreted here as parasequences belonging to these systems tracts.

A tectonic overprint may complicate the foregoing eustatic sequence-stratigraphic interpretation, given the tectonically-active basin setting. Fluctuations in subsidence rate, of both the lake floor and the sill, may have produced their own signature, compounded with the eustatic signal.

A detailed re-examination of the Ross and Bude Formations is warranted in order to refine the sequence stratigraphic interpretation, and to fully document the extent of wave influence.

Tides
No evidence for tides has been reported from the Ross or Bude Formation. Tides in Lake Bude would have been negligible when the lake was cut off from the sea (Talbot and Allen, 1996), potentially increasing as sea level overtopped the sill. Rare tidal sedimentary structures have been reported in Namurian and Westphalian deltaic cycles in northern England (Brettle *et al.*, 2002 and references therein), probably within TST and HST intervals when sea-connected lakes were temporarily re-established over the delta plains.

Flaggy siltstone facies
This rare facies, occurring at several levels in the Bude Formation (King, 1966), contains fish trails and crab trackways (Higgs, 1988, 1991). It has the coarsest grain size of the three "background" facies recognized by Higgs (1991), namely dark mudstone, light mudstone and flaggy siltstone (lumped here as "shales" in the first two facies of Table 1). Depths of no more than a few tens of metres, and fresh water, were proposed for the flaggy siltstone by Higgs (1988, 1991). Further studies are required to place this facies in its sequence stratigraphic context.

Water depth
A water-depth range of 5-100m is estimated here for those portions of the Ross and Bude Formations containing hummocks and HCS (the bulk of the formation, except for the 1-10m shale intervals containing the fossiliferous bands). The presence of HCS suggests deposition above storm wave base which is up to 200m deep in the ocean (Johnson and Baldwin, 1996) but is substantially less in lakes due to the reduced fetch. Similarly, the absence of nearshore facies suggests deposition below fairweather wave base, which is about 10m in the ocean (Johnson and Baldwin, 1996). During deposition of the fossiliferous shale bands representing glacioeustatic highs, the water depth is estimated to have increased to as much as 150m, based on the maximum likely amplitude of glacioeustatic sea-level fluctuations. This amplitude was approximately 150m (Reading and Levell, 1996, table 2.2), although only part of this amplitude would have overtopped the lake sill. In contrast, water depths of 300-500m were proposed for the Ross Formation by Martinsen *et al.* (2000, fig. 9), based on unspecified criteria.

Slumps and slides
The Ross-Gull Island and Crackington-Bude successions contain intervals, up to 20m thick, of contorted intraformational facies, interpreted as slumps and slides (Gill and Kuenen, 1957; Burne, 1995; Freshney *et al.*, 1979; Gill, 1979; Melvin, 1970, 1987; Strachan, 2002; Martinsen *et al.*, 2003). Slumps have erosively planed tops (Gill and Kuenen, 1957) and are capped by a discontinuous sandstone bed up to 50cm thick (Gill and Kuenen, 1957; Burne, 1970). Sand volcanoes can occur in various positions (Gill and Kuenen, 1957): (i) directly on top of the slump; (ii) on top of the capping sandstone bed; (iii) within the capping sandstone bed which, in some cases, appears to consist of merged volcanoes (Gill and Kuenen, 1957, plate xxxv, fig. 1); and (iv) on top of a shale bed overlying the slump (Gill and Kuenen, 1957, fig. 5). Volcanoes can have either concave-up flanks and wide craters, or convex-up flanks (which can truncate internal layering) and narrow craters (Gill and Kuenen, 1957).

Each slump or slide may have been triggered by an earthquake (Gill and Kuenen, 1957) or by a tsunami wave from a distant seismic event (Einsele *et al.*, 1996). Slumping and sliding are generally uncharacteristic of shelves, due to the generally low gradient. However in this case slumping and sliding may have been facilitated by the weak cohesion of freshwater muds, and/or by the presence of methane bubbles, which can be generated at much shallower depths (cm) in freshwater lake-floor muds than in brackish or marine muds (m) (Curtis, 1977). Microscopic plant fragments (Higgs, 1991) may have been the source of the methane. Slumping and sliding may also reflect temporary local increases in gradient on the Lake Bude shelf, caused either by fault-block tilt, produced gradually (pre-earthquake strain) or instantaneously (syn-earthquake), with the earthquake serving to trigger sediment movement; or by depositional relief generated by differential compaction of sand tongues relative to their muddy flanks. Erosional truncation of the slump top may have been due to waves or a unidirectional current (Gill and Kuenen, 1957). For example, a tsunami backflow current could be responsible for the erosion, followed by deposition of a discontinuous tsunami sandstone bed (Einsele *et al.*, 1996), filling minor relief on the slump. Sand volcanoes directly overlying the slump reflect dewatering of the slump immediately after it stopped moving. In contrast, volcanoes separated from the slump by a shale bed reflect a dewatering episode long after slumping, possibly caused by storm waves capable of liquefying sea-floor sediment (Owen, 1987). Storm waves could also be responsible for shaping the convex-up volcanoes, in a manner similar to wave-shaping of conventional storm-bed hummocks (P. Myrow and J. Southard, *field observations and pers. comms., 2003*).

Important in the interpreted shallow-water context is that slumping and sliding involved lateral translation of a sediment package (on a basal slip plane or shear zone) whose "emplacement" caused no shallowing, except at the distal ramp (cf. Martinsen, 1989, figs 4, 6). Groups of strata-bound extensional faults (*e.g.* Edmonds *et al.*, 1979, plate 9) may represent slide heads (Martinsen, 1989, fig. 4).

Sand-on-sand loading within sandstone packets (Melvin, 1986; Collinson *et al.*, 1991) may also reflect earthquakes or tsunamis. Associated liquefaction and water escape may have blurred pre-existing lamination (Owen, 1987).

Slurry beds

Slurry beds, which are a minor component of the Ross and Bude Formations, have been interpreted as mudflow- or liquefied-flow beds (Burne, 1970; Chapin *et al.*, 1994), and as seismites formed *in situ* (Higgs, 1991, 1998).

Lack of evidence for emergence

The Bude Formation is 1,290m thick (Freshney *et al.*, 1979), while the Ross is 460m thick. It is curious that such thick shallow-water successions lack nearshore (shoreface or delta-front) facies, or evidence for emergence (*e.g.* palaeosoils, ravinement lags, fluvial facies). This suggests that deposition was maintained within a narrow depth window on the shelf. Two regulating mechanisms, both mentioned above, are envisaged: a lake sill, which prevented eustatically forced emergence; and periodic wave-grading events, limiting shelf aggradation. Other periods of shallowng by aggradation (HSTs) were too brief to result in emergence. For example, given an initial estimated highstand water depth of 100-150m (*see above*), net depositional shallowing of only 20-40m can be calculated based on the following assumptions: (i) a low-order HST lasted for 20% of the duration of a 200,000-year Ross cycle, *i.e.* 40,000 years; (ii) during this time interval, one 0.2m storm bed and one 0.2m river-fed turbidite were deposited every 1,000 years (low frequency in brackish water; total is 16m); (iii) 1mm/yr mud deposition, totalling 40m, some of it eroded during turbidite and storm events; and (iv) subsidence is 12m (300m/Ma; *see above*).

The "sill-limited regression" model implies that the Clare and Bude shelf areas lay below sill level, and were too far offshore, even at lowstands, for unforced shoreline progradations to arrive before the next transgression (Higgs, 1987, 1991).

Gull Island Formation

This mudstone-dominated formation, which rests on the Ross Formation in County Clare, was interpreted as a slope deposit by Collinson *et al.* (1991), based partly on the abundance of slumped intervals. Martinsen *et al.* (2003) considered the lower part to consist of deep-water trough-axis deposits and the upper part of delta-slope deposits. A shelf interpretation is preferred here for the lower part, based on facies similarities with the Ross Formation, including small and large near-symmetrical ripples (Plates 9, 10). A shelf setting is consistent with gradation downward into the Ross Formation and upward into the deltaic deposits of the upper Gull Island Formation (Martinsen *et al.*, 2003) and Central Clare Group (Pulham, 1989). The lower sandstone-shale ratio, compared to the Ross Formation, is interpreted here to reflect a more distal setting, in an along-shelf sense (*i.e.* further from a sand-supplying river). Slumping accords with a tectonically-active, freshwater shelf setting, as discussed above.

Palaeocurrents

Palaeocurrents measured from sole marks in the upper Crackington-Bude succession (the entire Westphalian A-B interval in Fig. 1) show a gradual 180^{o} swing over time. They are directed towards the NE at the base, then to the SE, and then finally to the SW at the top (Freshney *et al.*, 1979, fig. 6). This palaeocurrent variation is conventionally attributed to sediment being supplied from three sides of the basin (*e.g.* Thomas, 1982, fig. 3.6); however, given the mineralogical and grain-size uniformity of the Crackington and Bude sandstones (Edmonds *et al.*, 1979; Freshney *et al.*, 1979; Melvin, 1986), this seems unlikely. An alternative interpretation of the flow reversal is that the northward drift of the British Isles across the palaeoequator (Scotese and McKerrow, 1990, figs. 17-19) caused a switch in geostrophic (Coriolis) deflection. Thus, river-fed turbidity currents on a presumed SE-facing shelf were initially deflected leftward while the shelf was in the southern hemisphere, resulting in NEward along-shelf flow; then they were deflected rightward in the northern hemisphere, causing SWward flow. These two phases were separated by a period of down-shelf flow during zero deflection at the equator. This model would be feasible given northward plate drift at a rate of 0.02m/yr, the average for the late Devonian to early Permian (calculated from Scotese and McKerrow, 1990, figs 16-20). For an interval of 11Ma (the duration of the Westphalian A-B in Fig. 1), drift would total 220km, or about 2^{o} of latitude. The distance may have been greater if short-term drift rates were faster or the duration was longer.

Comparisons with modern shore-parallel geostrophic bottom flows are appealing, although these are induced by storm winds (Walker and Plint, 1992) rather than river inflow. Along-shelf geostrophic flows have also been invoked in some ancient storm deposits (Myrow *et al.*, 2002 and references therein). In modern deep-sea turbidites, the influence of the Coriolis effect has been known for some decades, and

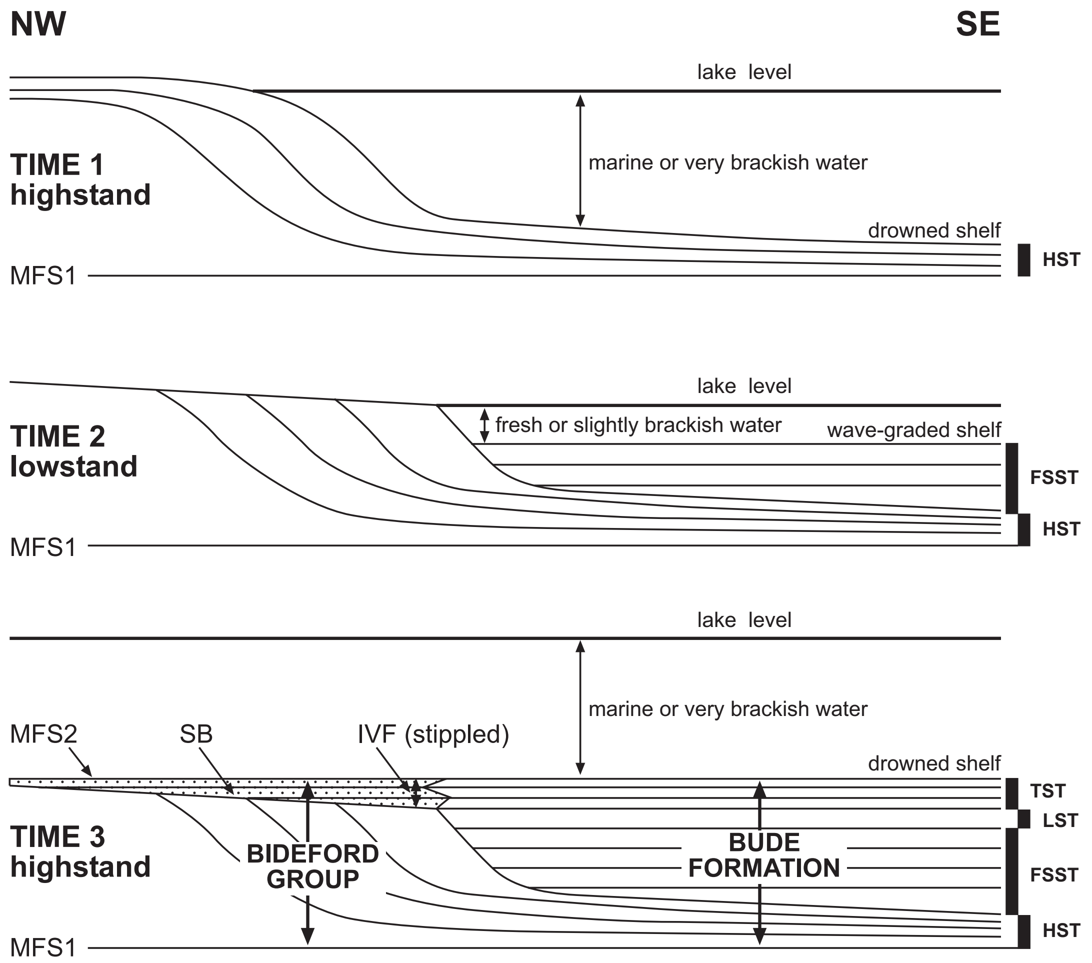

Fig. 3. Sequence stratigraphic model showing how Bude Formation aggradational lake-shelf deposits interweave laterally with essentially progradational delta cycles of the Bideford Group. Three stages of a complete lower-order glacioeustatic sequence are shown (*see text*). The Bude and Bideford successions are interpreted here as a stack of such sequences. The cross-section runs along the axis of an incised valley fill (IVF). FSST, HST, LST and TST are falling-stage-, highstand-, lowstand- and transgressive systems tracts; MFS is maximum flooding surface. Each Bude Formation FSST-LST pair is temporally equivalent to a sequence boundary (SB) at the base of a fluvial sandstone IVF, or its along-strike interfluve, in the Bideford Group.

for example causes inequality in channel-levee heights (Komar, 1969); but it has seldom been invoked in sedimentological studies of *ancient* turbidites.

The Coriolis force is proportional to the sine of latitude, so the effect is 20 times weaker at 2° than at 45° latitude. On the other hand, the force is infinitely stronger at 2° degrees than at 0°, where it is zero. Even at low latitudes, significant geostrophic effects due to Coriolis deflection can occur. For example, at 5° latitude, the Omo River plume entering Lake Turkana in Kenya is deflected alongshore by this mechanism (Yuretich, 1979; Talbot and Allen, 1996, p. 87).

The amount of deflection experienced by an unconfined underflow moving down a gradient depends on various parameters such as flow thickness and duration, and the magnitude of the Coriolis force (see Hill, 1984). The maximum theoretical deflection can be as much as 90°, resulting in along-slope flow, if the underflow is of sufficient duration. Flows lasting for days, such as those envisaged in the Ross and Bude Formations, are more likely to achieve maximum deflection than are slump-generated turbidity currents lasting only a few hours. The slow (dm/sec) Ross and Bude underflows would need less distance to be deflected than would faster (m/sec) slump-generated flows, since the turning radius is proportional to velocity (Hill, 1984).

Palaeocurrents in the Ross-Gull Island succession also swing gradually over time, through 90°, first towards the NE then the SE (Collinson *et al.*, 1991, p. 235 and fig. 8). An anomalous cluster of NW-flowing palaeocurrents near the base of the succession, recorded at one locality but not at others (Collinson *et al.*, 1991, fig. 8; Chapin *et al.*, 1994, fig. 5i-m), is of unknown significance. The swing is attributed here to the Coriolis effect decreasing to zero as the "Clare shelf", postulated to have faced SE, drifted northwards onto the equator. Clare reached the equator earlier

(Namurian) than Bude (Westphalian) because then, as now, Clare was slightly further north.

Superimposed on these long-term swings, palaeocurrents in the Ross Formation can vary substantially from one sandstone bed to the next, by as much as 80° (Collinson *et al.*, 1991, p. 235). Such variability may be a consequence of mixing readings from sole marks and from slightly asymmetrical ripples, the latter reflecting wave orientations rather than underflow direction.

The 90° discrepancy between the SE vergence of slump folds in the Gull Island Formation and NE-directed palaeocurrents, reported by Collinson *et al.* (1991), motivated their model of turbidites flowing *along* the foot of a prograding deep-water slope. However, this interpretation was rejected by Wignall and Best (2002, p. 540) as requiring "turbidity flows to have flowed along rather than down a slope; a physical impossibility".

LATERAL AND VERTICAL TRANSITION TO DELTAIC FACIES

Bude Formation cycles are primarily aggradational as they are dominated by combined FSST-LST deposits, which accumulated by lake-shelf aggradation rather than lake-margin progradation (*see above*). The intervening HSTs are inferred to grade shoreward into much thicker, progradational, delta-slope deposits of the Bideford Group, which crops out immediately north of the Bude Formation (de Raaf *et al.*, 1965; Elliott, 1976). The Bideford Group comprises nine deltaic cycles, each of which broadly coarsens upward. Correlation by fossiliferous bands shows that the Bideford Group is contemporaneous with the lower Bude and upper Crackington Formations (Eagar and Xu Li, 1993); the nature of the lateral contact (fault *versus* transitional) is uncertain due to poor exposure. According to Freshney and Taylor (1980, p. 384), "the massive sandstones of the Bude Formation (pass) into the distributary channel sandstones of the delta". In contrast, Reading (1965) presented faunal and sedimentological evidence which led him to suggest that the Bideford Group is bounded by north-vergent thrusts, an interpretation supported by reduced vitrinite reflectance values compared to those of neighbouring outcrops of the Crackington and Bude Formations (Cornford *et al.*, 1987), and consistent with the north-migrating foreland-basin setting (*e.g.* Gayer and Jones, 1989, fig. 2C). Nevertheless, prior to this structural deformation, the Bude and Bideford successions are inferred here to have interwoven laterally (Fig. 3). Each Bude FSST-LST correlates with a sequence boundary inferred to separate the Bideford delta cycles. The sequence boundary is overlain in some cycles by a thick sandstone unit (de Raaf *et al.*, 1965; Elliott, 1976), of kilometric lateral extent (Edmonds *et al.*, 1979), interpreted here as an incised-valley fill. Palaeocurrents in the Bideford Group indicate SEward delta progradation (de Raaf *et al.*, 1965), consistent with the SW-NE Crackington-Bude shelf orientation inferred above.

Similar coarsening-up deltaic cycles are present in the Central Clare Group (Fig. 1; Pulham, 1989), interspersed with sandstone units interpreted as incised-valley fills (Elliott and Pulham, 1991; Elliott *et al.*, 2000). The Central Clare Group gradationally overlies the upper Gull Island Formation delta-slope deposits and lower Gull Island shelf deposits *(see above)*.This large-scale, shelf-to-delta regressive transition, of R1 age (Fig. 1), may have been controlled eustatically or tectonically. A similar transition in SW England (base of Bideford Group: Fig. 1) is possibly of the same age (R1 or H according to Eagar and Xu Li, 1993).

SANDSTONE ARCHITECTURE: COMPARISON WITH DEEP-SEA FANS

The tabular sandstone packets which characterize the Ross Formation have been interpreted as deep-sea-fan lobes traversed by shallow channels (Chapin *et al.*, 1994; Martinsen *et al.*, 2000), and those in the Bude Formation have been similarly interpreted (Melvin, 1986, 1987; Burne, 1995, 1998). Instead, the packets are interpreted here as lake-shelf sandstone tongues, up to 10m thick, a few kilometres wide, and possibly 20-30km long. Ross-Bude sandstone architecture is interpreted in terms of stacked tongues connected vertically by small channels that cut the tongues in their proximal reaches (Fig. 4). In contrast, conventional deep-sea fans comprise offset fan lobes connected to an inner-fan channel (Richards and Bowman, 1998, figs 6-9). Ross-Bude sandstone tongues are likely to differ in the following ways from deep-sea-fan lobes:
(i) plan-view shape; (ii) lateral extent (width and length); (iii) rate of thinning along and across the sand-transport direction; (iv) rate of lateral grain-size change; (v) internal configuration (*e.g.* "bed lengths" of Chapin *et al.*, 1994); and (vi) degree of overlap of successive tongues. Fan lobes have no vertical connectivity (except indirectly via the inner-fan channel), unlike Ross-Bude tongues.

For these reasons, the Ross and Bude Formations are inappropriate as outcrop analogues for deep-sea-fan reservoirs.

PETROLEUM RESERVOIR POTENTIAL

The Ross and Bude Formations are potential oil or gas reservoir units in view of their great thickness (460m and 1,290m respectively), high sandstone

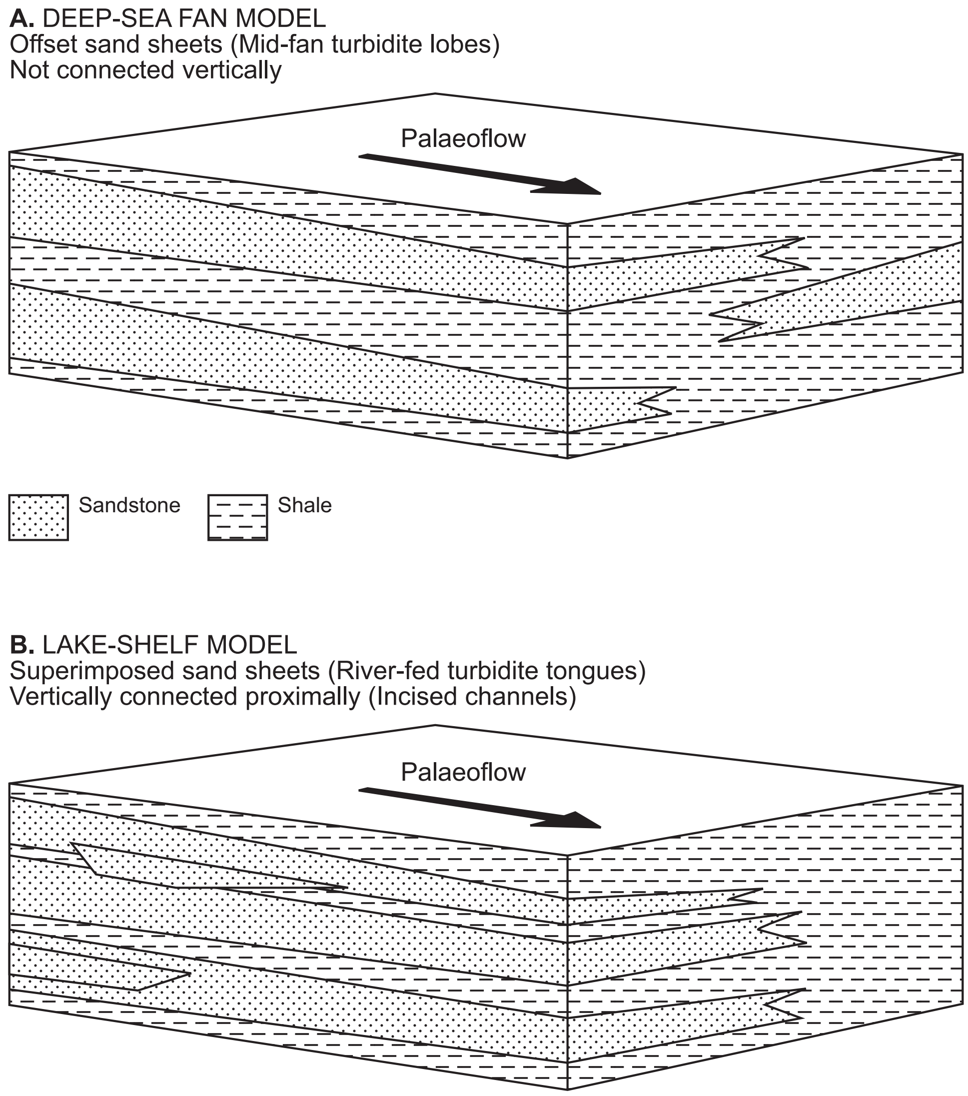

Fig. 4. Schematic sandstone architecture in: (A) the mid-fan area of a sand-mud rich deep-sea fan (cf. Richards and Bowman, 1998); and (B) a lake shelf receiving river-fed underflows. In (A), sandstone sheets are offset fan lobes, fed by an unseen inner-fan channel; they can overlap and are not vertically connected. In (B), sandstone sheets are river-fed tongues; they are essentially superimposed and can be vertically connected by channels incising the inner tongue. Channels are sinuous (*see text*) and "exit" the front-left panel obliquely. Note distal and lateral thinning of sheet sands in both cases. Not to scale.

content (c. 70%; Freshney *et al.*, 1979; Chapin *et al.*, 1994), and the presence of thick individual sand bodies (up to 10m) connected vertically by channels. A sand fairway corresponding to the Lake Bude shelf is predicted, beneath post-Variscan basin successions onshore east of Bude and offshore west of Ireland and SW England (Fig. 2). At outcrop, intergranular porosity in the quartzose, fine- and very-fine-grained sandstones of the Ross and Bude Formations appears to be generally low, estimated at less than 5%. However, porosity could reach 5-10% in the subsurface, because some of the cement is carbonate (Edmonds *et al.*, 1979; Freshney *et al.*, 1979), giving potential for secondary porosity by dissolution. Fracture porosity produced by Variscan folding and faulting may locally be well developed.

CONCLUSIONS

The Namurian Ross Formation of western Ireland is interpreted here as a shallow water lake-shelf deposit, rather than a deep-sea fan deposit. Key evidence supporting this new interpretation is: (i) a strong similarity with the near-coeval (Westphalian) Bude Formation of SW England, interpreted previously in terms of lake-shelf deposition; and (ii) the presence of wave-influenced sedimentary structures in the Ross Formation, reported here for the first time. The Ross Formation was deposited in the same foreland basin as the Bude Formation, as opposed to a rift basin confined to western Ireland. Sandstone architecture in the Ross and Bude Formations is likely to differ strongly from that of submarine fans.

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