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The Variscan Orogeny: the development and deformation of Devonian/Carboniferous basins in SW England and South Wales

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The upper Palaeozoic Orogenic Province of SW England is a part of a belt of Devonian and Carboniferous basins that extended from Devon and Cornwall through to Germany, some 800 km to the east. Their complex sequence of basin development and phases of deformation, described in this chapter cumulatively comprise the Variscan Orogeny in this region.

Synchronously with the Devonian events within the Variscan Orogen, the mainly fluvial facies of the Old Red Sandstone filled basins in the Avalonian continent north of the Variscan front (Chapter 6). During the succeeding Carboniferous, basins within the continent were mainly extensional in origin, until a period in the late Carboniferous when many basement faults were inverted (Chapter 7) resulting in uplift of the basin fill, that initiated a new palaeogeography at the start of the Permian.

The South Wales Basin represents a transitional zone between the mobile Variscan belt and the continent to the north. This transitional position is reflected in the Devonian by the interdigitationof the Old Red Sandstone facies and marine sediments at the northern margins of the Variscan basins (Chapter 6). Throughout the Dinantian and Namurian the succession within the South Wales basin had much in common with successions in basins within the continent to the north (Chapter 7). It was not until the Silesian that Variscan deformation affected basin development and caused its deformation (see this chapter).

The Variscan of SW England

The Variscan orogen in Britain and mainland Europe

The Upper Palaeozoic massif of SW England (Fig. 10.1), is situated on the northern margin and forms an integral part of the Variscan orogen of central and western Europe. The orogen is some 1000 km wide, and represents tectonic activity from the Early Palaeozoic to the Early Permian, and was in large part contemporaneous, and interactive with the Caledonides (e.g. Soper 1986*b*). Correlation of the province with the Rheinisches Shiefergebirge of Germany, some 800 km to the east, is established (e.g. Franke & Engel 1982; Holder & Leveridge 1986*a*) and together they make up a major part of the Rhenohercynian Zone. That is the northernmost of the palaeogeographical and tectonic zones proposed by Kossmat (1927), which, with the Saxothuringian and Moldanubian zones, constitute the orogenic belt in Germany and central Europe (Fig. 10.2).

Bard *et al*. (1980) and Matte (1986) proposed interpretation of Kossmat's zones in terms of plate tectonics. At that time postulated settings for SW England were diverse; an ensialic back arc basin (Floyd 1982), part of a trans-Europe oblique shear zone (Badham 1982), and an intracontinental fold belt (Matthews 1984). The Gramscatho flysch (Hendriks 1949) and associated Lizard ophiolite (Strong *et al*. 1975; Styles & Kirby 1980) in southern Cornwall were seen by Barnes & Andrews (1986) as representing a small intracontinental basin generated in an E–W strike–slip zone across northern Europe. In contrast Holder & Leveridge (1986*b*) interpreted those as products of the closure of a more extensive oceanized basin. Correlation and continuity of stratigraphies and structures between SW England and central Europe (Holder & Leveridge 1986*a*) appeared to enhance the concept of an E–W linear Europe-wide Rhenohercynian oceanized zone and collision belt (Franke 1989). By extrapolation of Kossmat's zones through western Europe into Iberia (Fig. 10.2), Franke (1989, 1992) and Matte *et al*. (1990) defined sublinear microplates and separating collision belts, with southerly subduction zones in the Rhenohercynian and Saxothuringian and northerly subduction beneath the Moldanubian. Dextral interplate movements and small oblique basins were reinvoked, however, for SW England by Holdsworth (1989). He speculated that the E–W 'Start–Perranporth Line' might represent a terrane boundary, between the 'Old Red Sandstone' continent to the north and an Armorican microplate to the south, with transtension actively controlling intervening Devonian basin evolution.

The Rhenohercynian Zone in continental Europe is now considered to lie at or near the southern margin of the Eastern Avalonia plate (Franke 2000), the southerly of the three plates,

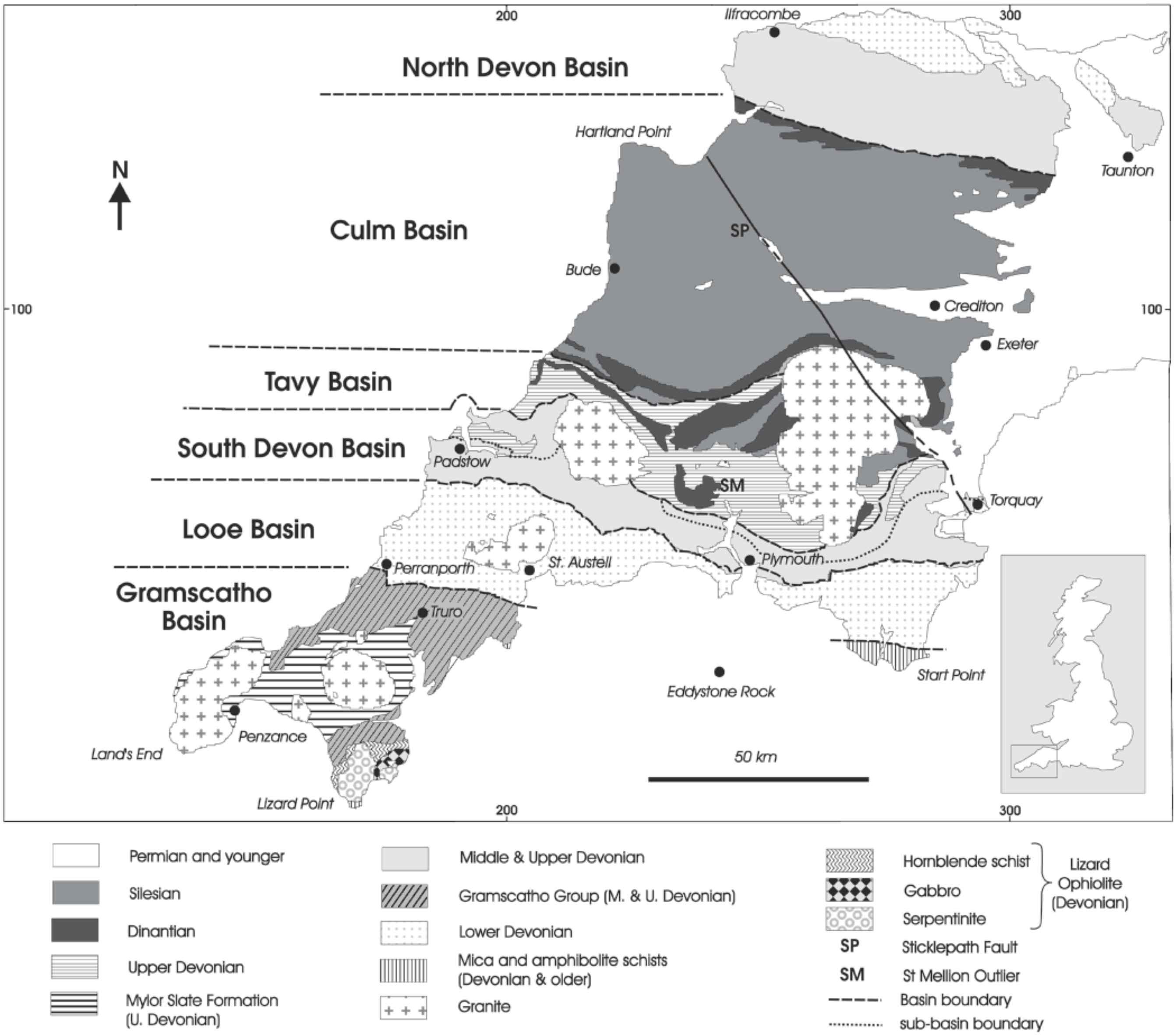


Fig. 10.1. Geological sketch map, based on BGS maps, showing the basins of SW England.

with Laurentia and Baltica, which collided to form the Caledonides (e.g. Soper 1986; McKerrow *et al.* 1991). Avalonia and Armorica were derived from Gondwana when in a southern peri-polar position in the Lower Palaeozoic, drifted north at differing rates (e.g. Trench & Torsvik 1991) and reassembled later in the Palaeozoic in equatorial regions. Armorica comprises an assemblage of continental microplates, with intervening marginal, basinal and ophiolite remnants, of the Saxothuringian and Moldanubian zones and their correlatives, which amalgamated during that process (Tait *et al.* 1997). The major Variscan oceanic separation, the Rheic Ocean, was proposed by Scotese *et al.* (1979), Cocks & Fortey (1982), and Van Der Voo (1983), to lie between the 'ORS' (Eastern Avalonia) continent and Armorica on the basis of palaeomagnetic and biostratigraphical evidence. More recent palaeomagnetic data have supported that view (see Tait *et al.* 1997). Recognition of an oceanic suture at the southern margin of the Rhenohercynian (Holder & Leveridge 1986*a*; Franke 1989) led some to propose that it was that of the Rheic Ocean (e.g. Oczlon 1993). The lack of a paired metamorphic belt, and apparently restricted related arc magmatism, indicated to Franke (1989) that it was a narrow ocean, up to 500 km being proposed by Franke *et al.* (1995). In Germany the ENE–WSE-trending Rhenohercynian suture lies between the Northern Phyllite Zone and the Mid German Crystalline High (of the Saxothuringian Zone) and within those tectonic units there are Ordovician and Silurian arc volcanics trending NE–SW. Those have been attributed to the closure of the Rheic Ocean (Franke & Onken 1995; Franke *et al.* 1995; Oncken 1997). Closure is thought to have been at the end of the Lower Palaeozoic (Tait 1999) immediately preceding Rhenohercynian rifting in the early Devonian along or close to the Rheic suture (Franke 2000).

McKerrow *et al.* (2000*a*) have accepted a close coincidence of the Rheic suture and the Rhenohercynian suture in SW England. Geological evidence confirming such coincidence, or the presence of a pre-Devonian Rheic suture immediately south of or within the province (beneath the Start–Perranporth zone with its locally strong secondary Variscan structures?) is awaited.

Devonian and Carboniferous basins of SW England

The Devonian and Carboniferous rocks of SW England are disposed within six juxtaposed E–W-trending basins (Fig. 10.1).

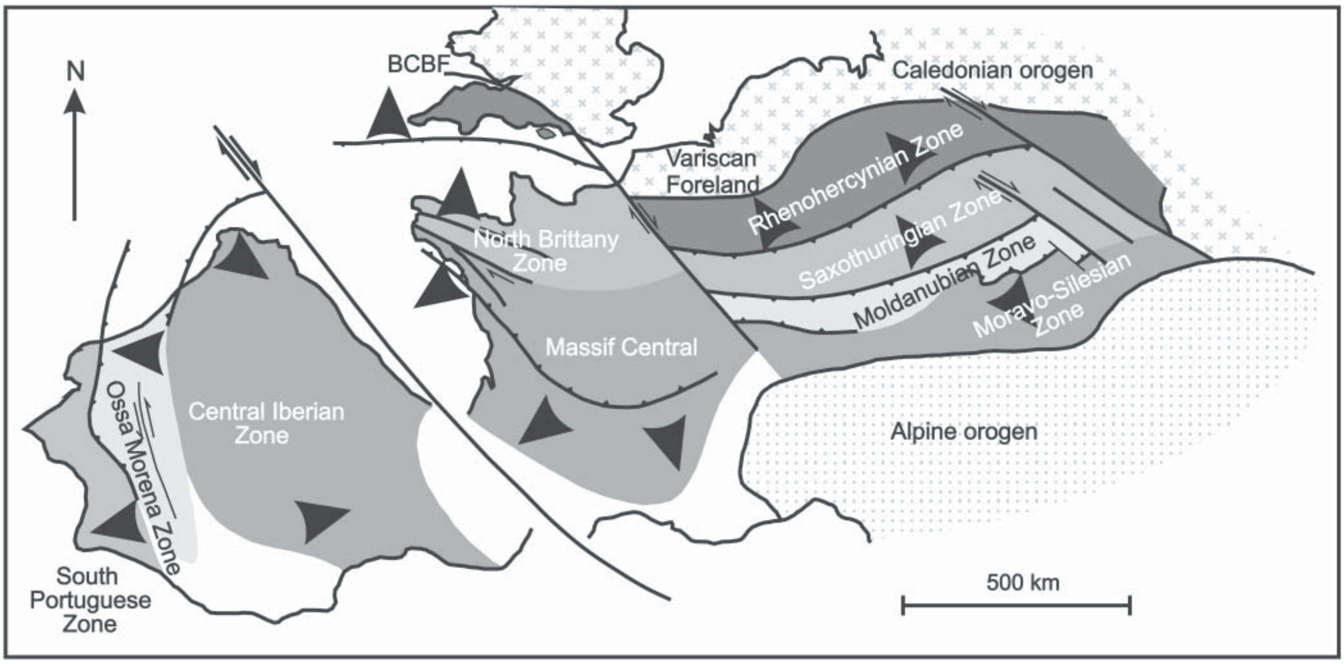


Fig. 10.2. European Variscan tectonic zonation, after Franke (1989), based on Kossmat (1927), as amended by Holder and Leveridge (1994). BCBF, Bristol Channel/Bray Fault Zone.

The evolution of those basins broadly reflects: (i) episodic Early Devonian–Dinantian continental rifting that developed a broad, complex, passive margin with oceanic lithosphere in the south; and (ii) Late Devonian continental collision and Carboniferous deformation of the passive margin.

In the southern part of the area now represented by the Gramscatho Basin and Lizard and Start complexes (Fig. 10.1), there was an intermediate stage. Here early rifting and the generation of oceanic lithosphere were succeeded by the formation of an active continental margin by the early Mid-Devonian. As a consequence, lithostratigraphical units here provide evidence of convergence-related deformation during Mid–Late Devonian times prior to continental collision.

Although the Gramscatho Basin formed during Early Devonian rifting, its infill (Figs 10.1 & 10.6) was largely derived during Mid–Late Devonian convergence, from its active southern continental margin. However, remnants of the earlier rift history, including oceanic lithosphere, are preserved. The Devonian–Dinantian infill of the other basins to the north (Figs 10.1, 10.8 & 10.10) reflect sedimentation and magmatic activity during the Early Devonian-Dinantian rifting alone. The Culm Basin initiated during this rift episode, but, following collision, and progressive basin closures and inversions to the south, it became a foreland basin in the Silesian.

This review emphasizes the role of extensional and convergent tectonics in determining the various stratigraphical successions of the province. The role of global sea-level changes due to eustatic, climatic and extraterrestrial events, and their possible bearing on Devonian successions of the province, have been discussed at length by House (e.g. 1983, 1992, 1996). The precise interrelationship or interplay between these factors within the province has yet to be determined.

Regional development

Devonian rifting and Mid–Late Devonian convergence

It has been recognized for some time that the Devonian–Dinantian lithostratigraphical divisions of SW England are compatible with the formation and development of a series of sedimentary basins during continental rifting (e.g. Matthews 1977). The Lizard ophiolite, in association with Gramscatho Basin tectonostratigraphy, has been interpreted as evidence that such rifting led to the oceanization of that basin (e.g. Barnes & Andrews 1986; Holder & Leveridge 1986*b*). The 'Basin and Rise' concept, imported to SW England from the Rheinishes Schiefergebirge of Germany (Goldring 1962), explained the common Mid–Late Devonian association of contemporaneous shallow-marine platform and basinal deposits between the Gramscatho and Culm basins (e.g. House & Selwood 1966; House 1975; Selwood *et al*. 1984). Most of these sequences were formerly assigned to the Trevone Basin (Matthews 1977) in the west or the South Devon Basin (Selwood & Durrance 1982) in the east (e.g. Selwood 1990). However, three major E–W Devonian basins, including previously unassigned Lower Devonian successions north of the Gramscatho Basin, are now identified (Leveridge *et al*. 2002); the Looe, South Devon and Tavy basins (Fig. 10.1).

The Gramscatho, Looe, South Devon and Tavy basins developed sequentially northwards from the Early Devonian, and rifting culminated with formation of the Dinantian Culm Basin. Pre-rift basement is not exposed, but is probably present at shallow depth locally (Leake *et al*. 1988). The timing of basin development is indicated and constrained by factors such as sediment supply blocking, and the availability of sediments derived from developing or isolated highs. The basins and associated highs reflect the formation of one or more half graben and/or graben and have their own characteristic stratigraphy, i.e. the province cannot be represented by a single 'layer-cake' model (Leveridge *et al*. 2003*b*).

The basins were variously interconnected, the Looe Basin being largely marine from the late Pragian and others essentially marine from the late Emsian (Fig. 10.3). Lithostratigraphical divisions of hemipelagic sedimentary rocks are recognized to form parts of more than one basinal succession. There was also along-strike variation of successions resulting from an overall westward deepening of the Looe and South Devon basins and greater submergence of intervening highs. All successions and part successions are fault bounded, dismembered or interdigitated by between one and three major episodes

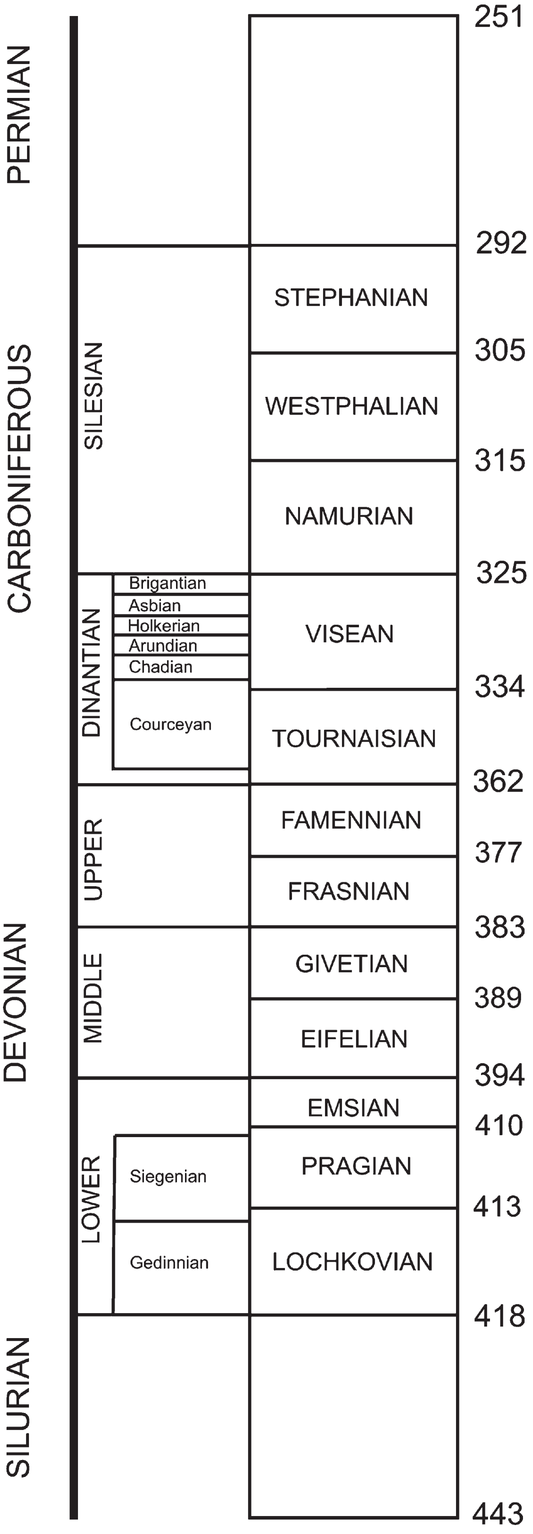


Fig. 10.3. Chronostratigraphic ages (Ma) of stages and series of Devonian and Carboniferous rocks, after McKerrow & Van Staal (2000), and of the Permo-Carboniferous boundary, after Menning *et al.* (2000).

of regional thrusting verging northwards and/or southwards. The relationship of principal lithostratigraphical divisions to this basin and high framework is indicated in Figure 10.8.

The connection of the apparently continuous North Devon 'Basin' succession to those of the contemporaneous rift basins to the south crop is not constrained, because of the crop of the intervening Culm Basin deposits, that link them only in the Dinantian. There are significant indicators however suggesting that they were indeed part of the same province. The major transgressions of north Devon, in the Late Emsian (?), Early Givetian and Late Famennian, show correspondence with the times of formation of the South Devon, Tavy and Culm basins respectively (see below, Fig. 10.11). This suggests that the passive margin subsided as a whole following major episodes of rifting and new basin formation to the south. There is also a direct link between basins in the Frasnian and Famennian. Purple and green sedimentary rocks are common to the neritic and continental facies of the Morte Slates and Pickwell Down Sandstones of the North Devon Basin and the Upper Devonian basinal facies of the South Devon and Tavy basins. Indeed, they extend into the deep-water Portscatho Formation sequence (Falmouth Series: Hill & MacAlister 1906) of the Gramscatho Basin.

Gramscatho Basin

Recognition that the association of peridotite, gabbro and mafic sheeted-dykes within the Lizard Complex in south Cornwall represented an obducted ophiolite (Strong *et al.* 1975; Bromley 1979; Kirby 1979; Styles & Kirby 1980) provided an important pointer to the setting of the province. In association with the dynamic stratigraphy of the Gramscatho Basin (Hendriks 1971; Holder & Leveridge 1986*b*) of south Cornwall, and its correlative lithostratigraphical development in Germany (e.g. Floyd *et al.* 1991), the significance of the ophiolite is enhanced in relation to the evolution of the Rhenohercynian.

Lizard Complex. The complex comprises the Lizard Ophiolite and representatives of possibly older basement (Fig. 10.4). Main components, (after Floyd *et al.* 1993*a*) are: (i) partially serpentinized peridotite; (ii) gabbros; (iii) sheeted basaltic dykes; (iv) meta-cumulates (Traboe Hornblende Schist); (v) mixed magma intrusions (Kennack Gneiss); (vi) metabasalt (Landewednack Hornblende Schist); (vii) metasediments (Old Lizard Head Series, OLHS); and (viii) orthogneiss (Man of War Gneiss, MOWG). The attribution of lithological units to tectonic units constituting the complex, the ophiolite, subjacent metamorphics or crystalline basement, remains a matter of debate, and indeed the credentials of the mafic rocks as an ophiolite have been questioned (Cook *et al.* 2002). The ophiolite, as generally understood, consists of (i)–(v) above, assigned to one (e.g. Leake & Styles 1984) or two (Bromley 1979) tectonic units, within which there are probably several thrust sheets (Jones 1997), with a structural thickness of a few hundred metres only (Styles & Kirby 1980; Rollin 1986). It rests upon a lower unit that has been considered to be (vi) and (vii) above, but may include (iv) (Power *et al.* 1996) or exclude (vi) as part of the ophiolite (e.g. Cook *et al.* 2002). Recent workers have agreed that locally the units structurally overlie the MOWG, part of a crystalline basement, forming inshore skerries off Lizard Point. The secondary nature of this faulted contact may, however, be indicated by associated semi-ductile/brittle fabrics (Jones 1997), and its obliquity to dated early Variscan fabric in the MOWG (Sandeman *et al.* 1997).

The peridotite comprises coarse lherzolitic, tremolite and dunite serpentinites. Separable stages of high-temperature and -pressure crystallization and recrystallization (Green 1964) and associated steeply inclined fabrics, mylonitic in part, orientated NNW–SSE, are attributed to progressive exhumation of mantle prior to obduction (Jones 1997; Cook *et al.* 2000). The Traboe Hornblende Schists are essentially metamorphosed cumulates, layered gabbros, dunites, pyroxenites and anorthosites, interpreted by Leake & Styles (1984) to be the first products of the ocean crust-forming process accompanying continental rifting. At Porthkerris these amphibolitic rocks are mylonitic, attributed to high-temperature ductile extension of the oceanic crust (Gibbons & Thompson 1991). A temporally associated plagiogranite vein has provided a constrained age of crust formation for the ophiolite (Clark *et al.* 1998*a*), which

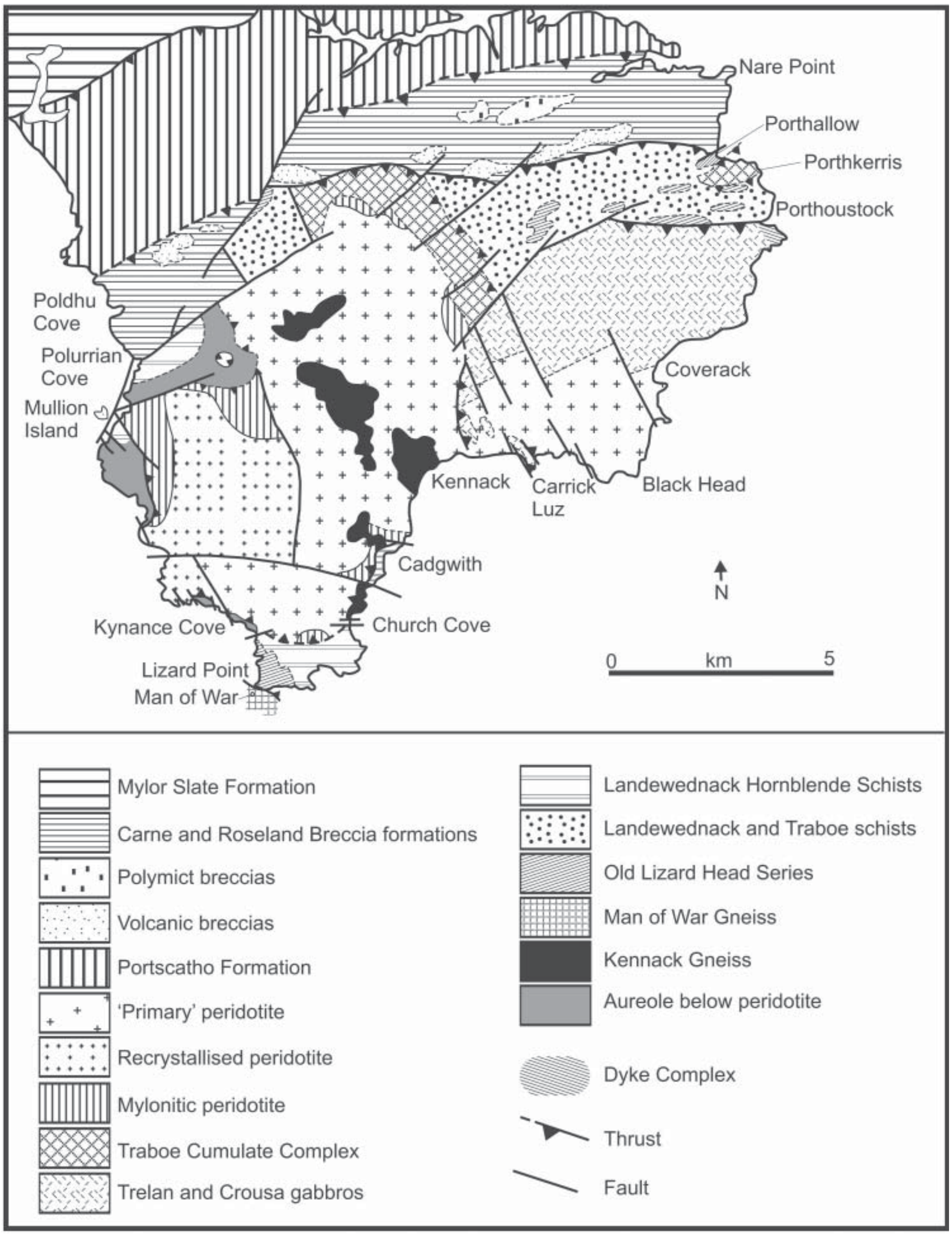


Fig. 10.4. Geological sketch map of the Lizard Complex, based on Floyd *et al.* (1993*a*) and Cooke *et al.* (2000).

is 397 ± 2 Ma (U–Pb single zircon), i.e. Emsian. The Crousa Gabbro, rather than superseding the mantle peridotite, intrudes it, and is cut by phases of basaltic dykes, trending NW–SE. The dykes, of N-MORB (MORB, mid-oceanic ridge basalt) composition, locally forming 80% of crop at Porthoustock, have been interpreted as a sheeted dyke swarm, as associated with a spreading centre (Kirby 1979). Amphibolite-grade metamorphism within these rocks, dated at approximately 390 Ma (U–Pb), has been attributed to their exhumation (Nutman *et al.* 2001; Cook *et al.* 2002).

The structurally lower unit of the complex comprises the OLHS and Landewednack Hornblende Schists to the south where they are in faulted contact (Jones 1997), and a similar association in the NE part of the complex (Floyd *et al.* 1993*a*; Power *et al.* 1996). The OLHS is largely metasedimetary quartz, mica and hornblende schists, locally with staurolite, kyanite and sillimanite (see Flett 1946), and the Landewednack Hornblende Schists, comprise meta-mafic volcanic and intrusive rocks of MORB affinities (Floyd 1976; Sandeman 1988). Designated by Floyd *et al.* (1993*a*) as sea-floor sediments and associated ocean floor basalts, or by Kirby (1979) as part of the ophiolite, an earlier history has been proposed for these rocks on the basis of early steep fabrics (e.g. Jones 1997; Cook *et al.* 2002). The OLHS is intruded by the granitoid Lizard Head Sill which has been dated at approximately 500 Ma (U–Pb) based on zircons interpreted by Nutman *et al.* (2001) to be primary magmatic, although zircons of similar age in the Landewednack and Traboe schists are interpreted as inherited. Later gently inclined mylonitic amphibolite facies metamorphic fabrics, with northwards transport indicators, within these rocks have yielded ^{40}Ar/^{39}Ar ages of *c.* 380 ± (av.5) Ma (Clark *et al.* 1998*b*) and slightly older (*c.* 390 Ma, with larger error margins) U–Pb ages (Nutman *et al.* 2001).

The Kennack Gneiss is a commingled magma intrusion at the thrust junction of the units, composed of E-MORB and crustally derived granitic melts. Intrusive and metamorphic age dates of 376.4 ± 1.4 Ma (U–Pb; Sandeman *et al.* 2000) and *c.* 370 Ma (Rb–Sr Styles & Rundle 1984; ^{40}Ar/^{39}Ar Sandeman *et al.* 1995), respectively, reflect its amphibolite facies metamorphism syn- and immediately post-emplacement (Sandeman *et al.* 2000).

The MOWG, composed of igneous rocks of gabbroic–tonalitic compositions, forms the undisputed base of the complex off Lizard Point. Geochronology by Sandeman *et al.* (1997) established that the gneiss, with an arc-like geochemistry, has a probable intrusive age (U–Pb, zircon) of about 500 Ma (Late Cambrian–earliest Ordovician), and a metamorphic cooling age (Ar^{40}/Ar^{39}, amphibole) of *c.* 374 Ma (Famennian) for amphibolite-grade recrystallization. This basement segment has Avalonian geochemical affinities (Sandeman *et al.* 1997) but its local status is uncertain. Possibilities are that it is part of the extensive nappe of basement mapped offshore (e.g. BGS 2000), a continental crustal block remnant of the Gramscatho Basin floor, or even a large olistolith within the underlying flysch nappe stack (see below).

The younger metamorphic dates within the lower tectonic unit, the Kennack Gneiss and the MOWG are generally attributed to Late Devonian thrust emplacement of the complex. The Kennack Gneiss represents magma comingling during that process, the felsic component derived by partial melting of basement or Gramscatho sediments, and mafic component derived from a mantle source similar to the ophiolitic rocks (Sandeman *et al.* 2000). Earlier dates within the ophiolite have been linked to mantle exhumation and uplift after Early Devonian oceanic crust formation and prior to obduction (Nutman *et al.* 2001).

Interpretations of the setting and generation of the oceanic lithosphere of the Lizard have focussed either on its internal fabrics and dyke orientations or the broader provincial and Rhenohercynian geology. Based on the former, a NNW–ESE spreading ridge axis (e.g. Gibbons & Thompson 1991), or small pull-apart basin (Badham 1982) linked to an E–W dextral intracontinental transform system (Barnes & Andrews 1986), or shear zone (Cook *et al.* 2002), have been proposed. In contrast basins to the north of the oceanized basin are E–W parallel, rather than oblique, within the province (see below). Available structural and stratigraphical data point to contemporaneous and ongoing pure shear extension, with formation within the province of E–W parallel, rather than oblique basins, in the Devonian and Dinantian (e.g. Leveridge *et al.* 2002). Early Devonian (Pragian) marine sediments (e.g. Meadfoot Group) straddle the 'Start–Perranporth Line' (Holdsworth 1989) and evidence for penecontempoaraneous independent Devonian basin evolution across it remains only speculative. The Lizard ophiolite was a small part of the Rhenohercynian ocean floor represented by MOR basalts associated with flysch nappes of Mid and Late Devonian deposits derived from its floor over 1000 km along strike (Floyd 1984, 1995; Holder & Leveridge 1986*a*; Franke 1989). A corollary of ophiolite obduction, as recognized by Coleman (1977), is the limited part of ocean floor evolution that it might represent, in terms of typicality, time and location of formation, and exhumation history. The interpreted strike-slip basin origin of the Lizard ophiolite and the orthogonal extensional rift origin of basins to the north, have yet to be satisfactorily reconciled. The possibility of ophiolite rotation during obduction is not fully constrained (e.g. Hailwood *et al.* 1984), and the consequences of its formation on the southern side of the basin at the time of transition from extensional to contractional regimes (see below) have yet to be addressed.

Gramscatho Basin, the rift phase. The main deposits of the Gramscatho Basin (Figs 10.1, 10.5 & 10.6), were interpreted as products of its active southern margin during basin closure and ophiolite obduction by Barnes & Andrews (1986), Holder & Leveridge (1986*b*) and Leveridge *et al.* (1990). MORB with deep-water sediments within the thrust stack also point to the oceanization of the basin, and Shail (1992) proposed that elements of both allochthonous and parautochonous sequences of the basin are relateable to the earlier rift phase of basin evolution.

The oldest biostratigraphically dated allochthonous rocks of the Gramscatho Basin are the early Mid–Late Eifelian (Sadler 1973; Leveridge 1974) deposits of the Pendower Formation (Figs 10.5 and 10.6) at the base of the Veryan Nappe. The formation is a hemipelagic shielded basin sequence of grey, green and brown slaty metalliferous (Mn, Cu, Ni, V, Zn) mudstones and interbedded radiolarian chert, with turbiditic limestone of pelagic platform provenance and subordinate thin beds of coarse lithic greywacke sandstone (Holder & Leveridge 1986*b*; Shail 1992). Near its lower thrust boundary pillowed basalt, LIL-enriched MORB at Tubbs Mill (Floyd 1984), interdigitates with the sedimentary rocks, and in sequence above the formation is the southerly sourced flysch of the Carne Formation (see below). Whilst attributing the MORB, chert and metalliferous mudstone association to the proximality of a spreading centre, the Pendower calciclastics and siliciclastics have been described by Shail (1992) as synrift deposits, sourced by tectonically controlled erosion of a within-basin remnant block of continental basement, capped by pelagic carbonate sediments. Deposition was interpreted to be within a northerly sub-basin, with its incipient Eifelian oceanic crust formation, of an asymmetrical Gramscatho Basin, the earlier (Emsian) Lizard ophiolite flooring part of a dominant southern sub-basin.

The parautochthonous succession (Holder & Leveridge 1986*b*) was probably deposited on the thinned northern continental passive margin of the basin. This is reflected in the presence in its uppermost formation, the Mylor Slate Formation (Figs 10.5 & 10.6), of contemporaneous intraplate tholeiitic basaltic rocks (Floyd & Al-Samman 1980). Locally, included tectonized granitoid xenoliths with model ages of 625 Ma and 851 Ma (T^{Nd}) have been interpreted as representatives of continental crystalline basement (Goode & Merriman 1987; Goode *et al.* 1987). Shail (1992) also proposed that some deposits of the underlying succession, the Porthtowan and Grampound formations (Fig. 10.5), were synrift sediments derived from the northern margin of the basin. Those included some of the thicker and coarser sandstone beds towards the top of the Porthtowan Formation and the Treworgans Sandstone Member of the Grampound Formation (Fig. 10.6). Constraints on the interpretation of northerly sourcing are the age (older than Frasnian but not known in detail), provenance (lithic greywackes compared to the few quartzo-feldspathic sandstones in passive margin sequences), and pathways of the sediments across the rifting margin to the north (see below).

Gramscatho Basin, the convergent phase. The infill of the basin was dominated by sedimentation generated along its active southern margin. Obduction of the Lizard ophiolite from the oceanized floor of the Gramscatho Basin during its closure has been proposed or modelled by several authors (e.g. Sanderson 1984; Barnes & Andrews 1986; Holder & Leveridge 1986*b*). Currently considered to have a late Early Devonian generation age (Clark *et al.* 1998*a*) it is located within a family of northward-verging thrust nappes (Fig. 10.5; Leveridge *et al.* 1984, 1990). These are major gently to moderately inclined structures not only mapped onshore but also traceable offshore in the South West Approaches Traverse (BIRPS & ECORS 1986) and commercial seismic profiles (e.g. Hillis & Chapman 1992) across some 300 km of the continental shelf.

Within the sedimentary rocks of the nappe pile, dip and younging are to the SE. The Carrick Nappe (Fig. 10.6), at the base of the allochthon (Holder & Leveridge 1986*b*: Leveridge *et al.* 1990), has a structural thickness up to 5.4 km onshore. It consists of the Portscatho Formation that is essentially Frasnian in age, but is possibly older in part, and probably extends up into the Famennian (Wilkinson & Knight 1989). This flysch facies sequence, of interbedded greywacke sandstone turbidites and dark grey mudstone, demonstrates progradation from outer fan to mid-fan depositional regimes. The overlying Veryan Nappe (Fig. 10.6), comprises the

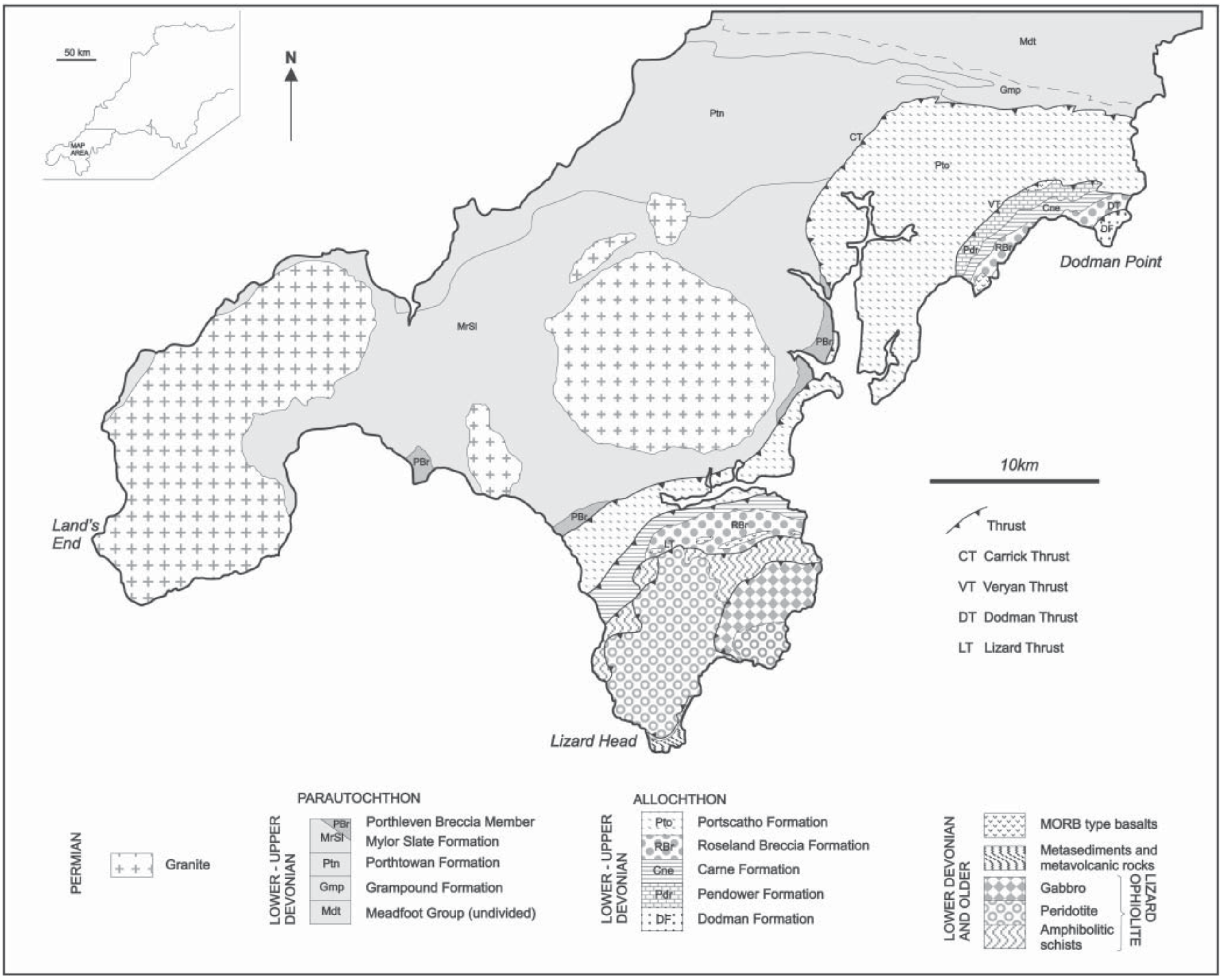


Fig. 10.5. Sketch map of the tectonostratigraphical divisions of south Cornwall, based on BGS maps, after Leveridge *et al.* (1990).

Pendower, Carne and Roseland Breccia formations, a succession some 2.6 km thick, ranging from the early Mid Eifelian–Givetian (Sadler 1973; Leveridge 1974), and probably early Frasnian (Hendriks *et al.* 1971). The latter attribution was unaffected by the subsequent revision upwards of the Givetian–Frasnian boundary by the International Subcommission on Devonian Stratigraphy (see Ziegler & Klapper 1985), as it was based on the conodont *Ancyrodella buckeyensis* (Stauffer) which is still assigned to the Frasnian (Over & Rhodes 2000). The Pendower Formation (see above) at the base of the nappe is succeeeded by the Carne Formation, a sequence of greywacke sandstone turbidites, olistostrome, channel-fill sandstone, slumped sandstone, and interlaminated mudstone, siltstone and sandstone. It is interpreted as an association of middle and upper fan distributary and interchannel deposits, slope deposits and proximal turbidites. The Roseland (and Meneage) Breccia Formation is a major olistostrome (e.g. Barnes 1983), composed of grey silty mudstone with dispersed clasts of sedimentary, metamorphic, volcanic and magmatic rocks, with interbedded sedimentary rocks, including breccias and conglomerates, and acid and basic volcanic rocks. Extrabasinal clasts, introduced by slumping, slides and sediment gravity flows, include shelf sediments of Lower Devonian, Silurian and Ordovician age, and a variety of schists and granitoids. Contemporaneous basic and acidic lavas, volcanic breccias and minor intrusions have MOR and within plate, calc alkaline, affinities (Barreiro 1996), and some larger clasts are ocean floor basalts (e.g. Nare Head, Roseland: Barnes 1983; Floyd 1984). Within this nappe sequence there is dramatic progradation from shielded basin sediments to climactic slope deposition. The overlying Dodman Nappe (Fig. 10.6), comprises an undated succession (Barnes 1983) of interbedded greywacke sandstone and mudstone. A reverse metamorphic gradient, from epizone to higher greenschist-grade mica schist, is present onshore in the Dodman Formation. Offshore data suggests that the Start Schists are part of this nappe. Hornblendic schists associated with mica schists at Start, which are also of flyschoid appearance, have N-MORB signatures (Floyd *et al.* 1993*b*; Merriman *et al.* 2000). Comparability with some schists of the lower tectonic unit beneath the Lizard ophiolite cannot yet be excluded (see above). A further nappe of crystalline rocks, the Normannian Nappe (Holder & Leveridge 1986*b*), is mapped offshore (SWAT 9: BIRPS & ECORS 1986; Edwards *et al.* 1989). It has surface expression in the garnetiferous gneiss of Eddystone Rock in Plymouth Bay, probably the Man of War Gneiss off Old Lizard Head and possibly the Old Lizard Head Series (Nutman *et al.* 2001). Locally interposed between the Dodman and Normannian nappes are the tectonic units of the Lizard Complex.

The flysch source, defined by sandstone-framework modal grains and their geochemistry, was a dissected continental margin magmatic arc (Floyd & Leveridge 1987), essentially of

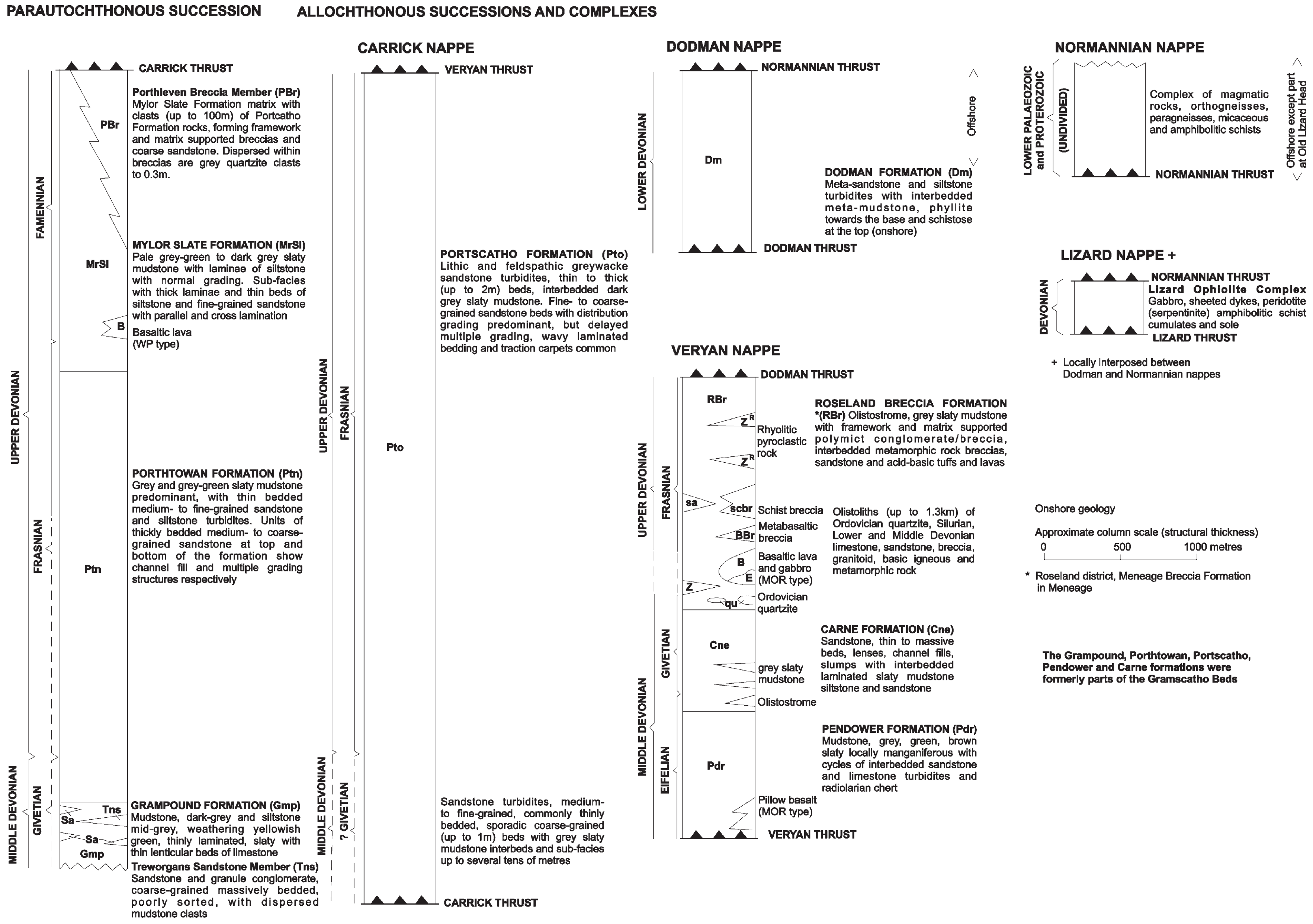


Fig. 10.6. Nappe and parautochthon successions in south Cornwall, based on BGS (1990) and BGS (2000).

pre-Devonian origin (Floyd *et al.* 1991). The extrabasinal clasts, including Lower Palaeozoic and Lower Devonian shelf sediments, schists and granitoids, are comparable to basement and Palaeozoic sequences in Armorica, and olistolith faunas also have southern affinities (Bohemian, Hendriks 1937; Armorican, Sadler 1973). A potential source between Cornwall and Brittany was identified by Ziegler (1982) as the Normannian High, an extension of the Mid German Crystalline High.

The thrust nappe stack (Fig. 10.5) has a transgressive ramped contact with the parautochthon. To the north of the Lizard it rests in a pseudoconformable manner on thick (up to 1.3 km) sedimentary breccias (Porthleven Breccia Member, Mylor Slate Formation), of Famennian age, at the top of the parautochthonous sequence (Leveridge & Holder 1985). The breccia includes clasts of the allochthonous Portscatho Formation showing pre-incorporation early (D_1) structures, in a matrix also deformed by D_1. The Mylor Slate Formation comprises mainly interlaminated and thinly interbedded dark grey mudstone, graded and locally cross-laminated siltstone and fine-grained sandstone. It is interpreted as representing, in its lower part (Shail 1989), an association of rise-slope proximal channel deposits and low-density turbidites and hemipelagites, and in its main part as distal turbidites (Leveridge *et al.* 1990). It overlies the Porthtowan Formation, an estimated 2.8 km-thick sequence of grey and grey-green mudstone with subordinate greywacke sandstone, part of the Gramscatho Group. It correlates in large part with the Portscatho Formation, but generally finer grain sizes and turbidite sedimentary characteristics are consistent with more distal deposition. The lower boundary is conformable with the Grampound Formation (Fig. 10.5), comprising coarse-grained sandstone grain-flow beds (up to 6 m) interdigitating with mudstone and wispy laminated siltstone and sandstone. The age and relationship of this formation and the Meadfoot Group (see below) immediately to the north has yet to be established. Significant however is that Meadfoot Group marine shelf rocks (Pragian–Emsian) extend from the north to the south of the 'Start Perranporth Line', suggesting that it did not have a significant influence on development of the basin, in terms of its constraint (see Holdsworth 1989) or sediment sourcing (Shail 1992).

Early, D_1, deformation in the flysch nappes is characterized by slaty cleavage and tight to isoclinal sheath folds that record NNW overriding translation (Leveridge *et al.* 1984). Dodson & Rex (1971) dated pervasive low-grade greenschist metamorphism (Warr *et al.* 1991) as a late Devonian–early Carboniferous event (K–Ar 370–350 Ma), but modern data (Clark *et al.* 1998*b*) extends that event back to the mid-Devonian (^{40}Ar/^{39}Ar 385 ± 2 Ma) in the Dodman Formation. Radiometric determinations show a broad pattern of younging uplift metamorphic ages, late mid-Devonian–early Carboniferous (*c.* 385–355 Ma), from SSE to NNW through the nappe pile.

The close temporal relationship of climactic sedimentation and deformation in south Cornwall indicates causal connection. The Devonian–earliest Carboniferous evolution of the dynamic stratigraphy of south Cornwall was modelled by Leveridge *et al.* (1990) by integration of nappe stratigraphies and the timing of deformation (Fig. 10.7). Although supported by Clarke *et al.* (1998*b*) and Sandeman *et al.* (2000) the orthogonal extent of the ocean and also the internal basin organization implied by that sequence diagram have been questioned (see above). However, there is concurrence (see Shail 1992) on the link between flysch sedimentation, northward tectonic migration, obduction of the Lizard ophiolite, and closure of the Gramscatho Basin. Those processes began with flysch deposition at the southern margin of the basin, sourced by the Normannian High, probably late in the Early Devonian, a timing suggested by the in-sequence nature of the onshore nappes. Northerly overthrusting of the Normannian Nappe, accompanied by sedimentation at its leading edge, initiated southerly subduction of the ocean floor. Detachment of the Dodman and Lizard nappes, inundation of the Pendower hemipelagic environment, and probable filling of the oceanized basin with flysch extending onto the northern passive margin of the basin, took place in the Mid Devonian. During the Frasnian forward propagation transferred motion to the new Veryan Thrust, carrying the Veryan Nappe onto more distal flysch. Further forward propagation took place in the late Famennian–early Carboniferous when the major Carrick Nappe was thrust out of the basin. Erosion produced a thick olistostrome at the thrust front and this was overridden by continuing nappe movement. Sandstone clasts within the olistostrome showing pre-incorporation deformation illustrate the penecontemporaneity of sedimentation and deformation within the migrating nappe pile. The emergence of the nappe onto the marginal sequence effectively marked the closure of the oceanized Gramscatho Basin, a process that apparently took some 35 Ma (Fig. 10.7), and the onset of continental collision.

Looe Basin

The southernmost of the major basins on the passive margin of the Gramscatho Basin is the Looe Basin (Figs 10.1 & 10.9), whose succession illustrates clearly the basin forming and filling processes. Within the basin there is in excess of 5 km of Lower Devonian and early Middle Devonian sedimentary rocks. These include the major lithostratigraphical divisions of the Lower Devonian, identified and mapped across the province by Ussher (e.g. 1890, 1903, 1907), the Dartmouth Slates, Meadfoot Beds and Staddon Grits. These have since become known as the Dartmouth Group, and the Bovisand and Staddon formations of the succeeding Meadfoot Group, respectively. Completing the basin succession is dark grey mudstone, of latest Emsian and Eifelian age (e.g. Jennycliff 'Slates' division of the Saltash Formation at Plymouth), attributed to the variously named formations that persist through the Devonian (Fig. 10.8). Commonly this upper part of the Looe Basin succession is occluded by overthrust Staddon Formation.

The Dartmouth Group, locally subdivided into formations (Dineley 1966; Seago & Chapman 1988), comprises terrestrial deposits in which there is a restricted non-marine fauna and fish remains. The latter include such as *Rhinopteraspis dunensis* and *Althaspis leachi* (White 1956; Forey in Ivimey-Cook 1992) that are indicative of the early Mid Siegenian (Pragian). As they are found at higher levels in the group, it may extend down into the Lochkovian (cf. House & Selwood 1966). Considered by Dineley (1966) to represent stable coastal mudflat regime, with lagoonal and fluviatile deposits, rocks of the group are now interpreted as the products of a more dynamic setting. To the east, the deposits of perennial lakes are interspersed with fluvial deposits (Smith & Humphreys 1989, 1991), whereas to the west the lacustrine regime was semi-permanent (Jones 1992; Leveridge *et al.* 2002). There in a sequence up to 3.3 km thick, with no base exposed, red and green mudstone and siltstone, deposited from suspension, constitute the lacustrine facies. Interbedded siltstone and sandstone, as laminae, lenses, sheet-beds, and amalgamated units, represent distal and more proximal deposition from fluvial-fed traction currents and underflows. Regularly developed quartzite interbeds indicate deposition of fluvial sediment from confined channel flow. Through the sequence pebbly mudstone units, with clast dispersion and composition indicative of mass-flow emplacement and intrabasinal derivation, appear to be the product of slope collapse. These features, together with growth faults and associated penecontemporaneous microgabbroic intrusions, point to rapid subsidence and instability of the basin. Cyclic sequences in the upper parts of the group include grey marine mudstone within the lacustrine and fluvial sediments, indicating hydrological continuity between the lacustrine regime and open marine conditions during periodic downfaulting. Contemporaneous volcanic rocks, predominantly basic lavas and tuffs, are alkaline and calc-alkaline basaltic rocks typical of a rifting regime with crustal contamination at an early stage of the process (Merriman *et al.* 2000).

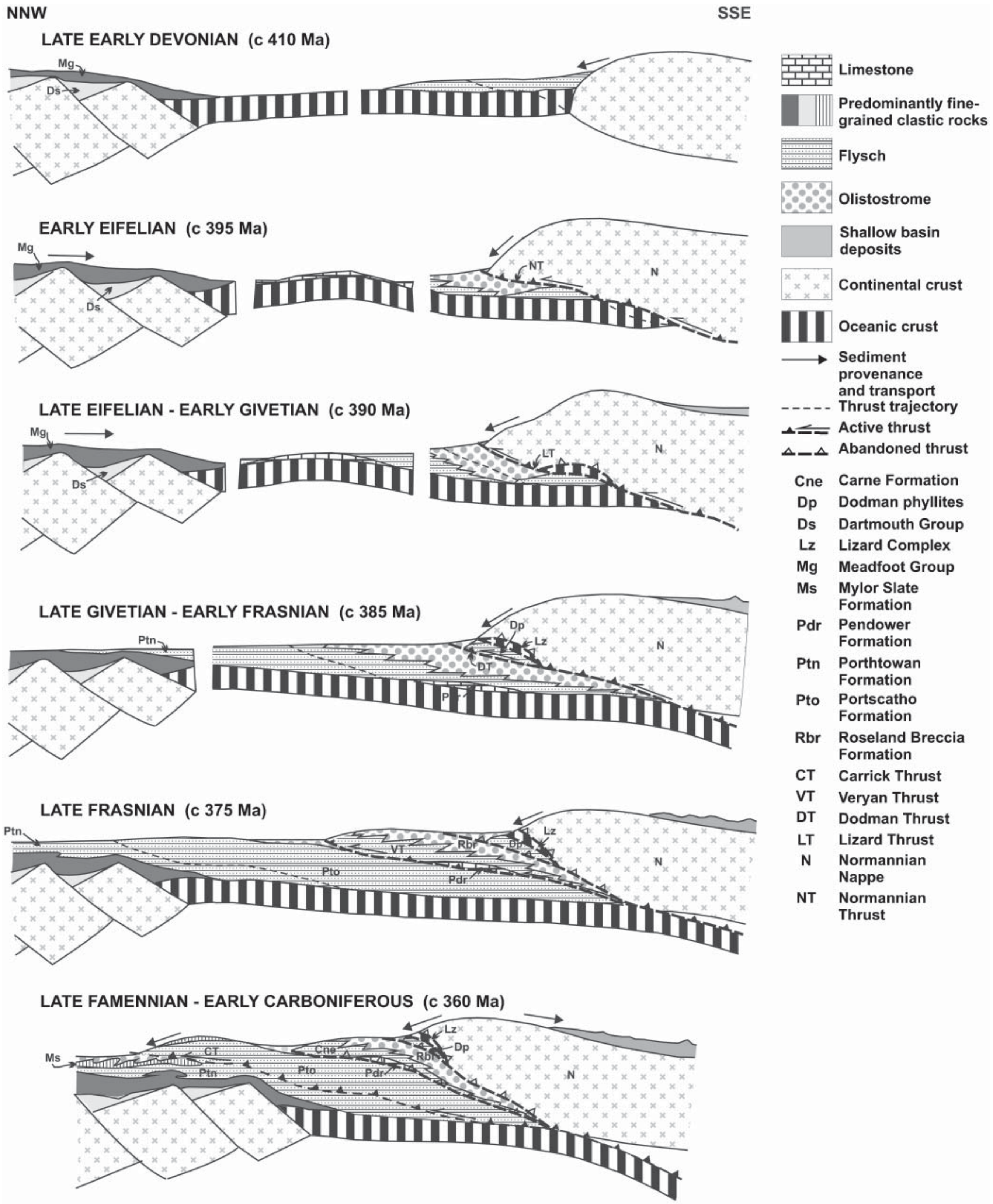
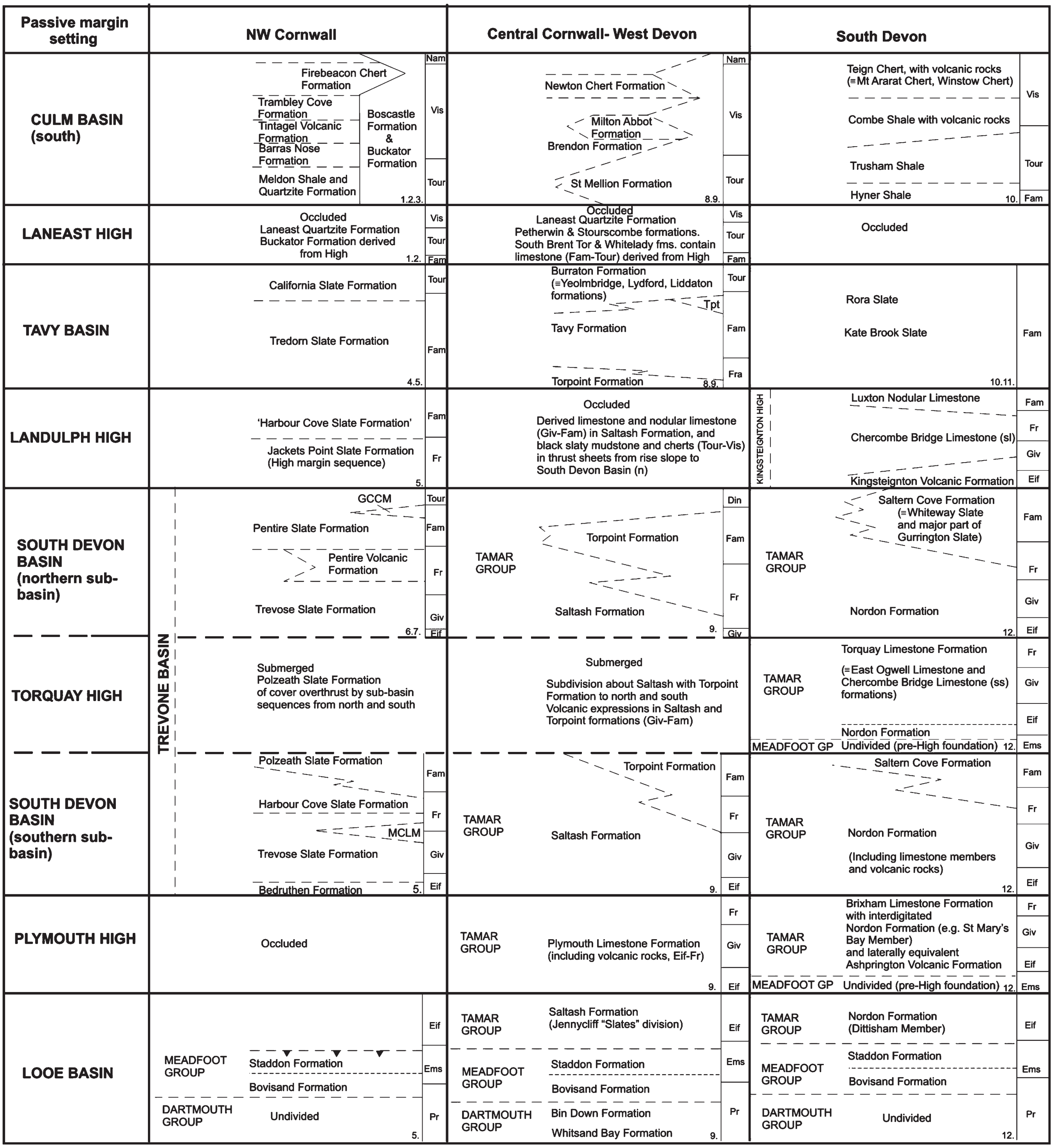


Fig. 10.7. Time sequence diagram indicating development of tectonostratigraphical divisions of south Cornwall, accommodating closure of the Gramscatho Basin, after Leveridge *et al.* (1990).

The Meadfoot Group represents establishment of the marine environment, a conformable base reflecting rapid inundation. Although extending locally onto the northerly marginal platform (Leveridge *et al.* 2003*a*) the group is largely confined within the Looe Basin where it comprises the Bovisand Formation and Staddon Formation (Harwood 1976; Seago & Chapman 1988).

The Bovisand Formation is a Pragian–mid-Emsian sequence representing a variety of offshore environments with sporadic lower shoreface reworking. Sandstone dominant members



⁺The successions/part successions are fault bound, dismembered/interdigitated by one to three episodes of thrusting verging northwards and/or southwards.

1-12 References for detail/review
1. Freshney et al. (1972) 2. McKeown et al. (1973) 3. Selwood et al. (1985) 4. Selwood and Thomas (1993) 5. Selwood et al. (1998) 6. Gauss and House (1972) 7. Austin et al. (1992)
8. BGS (1994) 9. Leveridge et al. (2002) 10. IGS (1976) 11. Selwood et al. (1984) 12. Leveridge et al. (2003a)

GCCM: Gravel Caverns Conglomerate Member MCLM: Marble Cliff Limestone Member Tpt: Torpoint Formation Submerged: Intrabasinal High without separable succession

Fig. 10.8. Devonian–Dinantian successions of the passive margin, in relation to its basin and high framework.

(up to *c*. 350 m), comprising cyclic sequences ranging from offshore-shelf facies to wave dominated and storm-dominated shoreline facies associations, reflect numerous phases of gradual progradation and rapid transgression.

The late Emsian Staddon Formation, with a thickness of 400 m (Selwood *et al.* 1998) or more, represents a major phase of such southward progradation. Its component fine- to coarse-grained sandstone beds record the transition from shallow-marine setting to high-energy regime. Deposition in shallow-marine embayments, with storm-wave-generated bars (Pound 1983), is succeeded by fluvial channel, levee and overbank deposition in a non-marine setting (Humphreys & Smith 1988), and the establishment of a substantial fluvial coastal sand plain.

Completing the Looe Basin succession is the lowest part of a predominantly grey slaty mudstone sequence that forms the marine background sedimentation of the passive margin southern crop throughout the Devonian. Variously termed the Trevose Slate Formation (see Selwood *et al.* 1998), the Nordon Formation (Selwood *et al.* 1984) and Saltash Formation (Leveridge *et al.* 2002), the main expression is within the South Devon Basin (Fig. 10.3). About Plymouth Sound the mudstone sequence, is the Jennycliff ('Slates') division of the Saltash Formation (Leveridge *et al.* 2002), which on the basis of conodonts (Orchard 1977, 1978; Dean 1994) and palynomorphs (Molyneux 1990; Dean 1992) ranges from late Emsian to late Eifelian. Abundant *Spirophyton*, with gutter casting and hummocky cross-stratification of sandstones low in the division are compatible with encroachment of the sea over the fluvial Staddon Formation and the recycling of sand from the drowned coastal plain by storm activity. In the upper mud-rich part of the sequence are interbeds of limestone and units of hyaloclastitic basalt lava and bedded tuff. Features of the limestone are indicative of deposition from high-energy currents with minimal transport of crinoid, brachiopod, coral, stromatoporoid and bryozoan debris from a shallow source.

The Looe Basin is interpreted as having developed on the extending and subsiding continental margin of the oceanized Gramscatho Basin. Initially terrestrial (Fig. 10.9a), the onset of marine conditions in the late Pragian indicates that the rate of subsidence of the basin exceeded sediment supply. The polycyclic shoreline developments of the Bovisand Formation represent episodic (sub)basin infilling and subsidence. Delta-top and fluvial sandstone of the Staddon Formation show that, in late Emsian times, deposition at the northern edge of the Looe Basin had outstripped subsidence. Rivers were feeding siliciclastic sediment into the basin across a piedmont from a source area to the north (Leveridge *et al.* 2002). With further regional extension the South Devon Basin developed to the north, isolating the Plymouth High between the basins, and shielding the Looe Basin from coarse clastic sediments. Related subsidence in the Looe Basin produced the starved basin facies at the top of its succession, interrupted by early Eifelian carbonate storm deposits derived from the shallows developing above the nascent Plymouth High.

South Devon Basin

The South Devon Basin (Figs 10.1 & 10.9) is composite, with northern and southern sub-basins. These are separated to the east about Torbay (Leveridge *et al.* 2003*a*) by a subdividing high, which is recognized by lithostratigraphical segregation near Plymouth, and in the west where it is also reflected in structural facing confrontation near Polzeath. The basin shows an overall deepening from east to west. Basin formation by half-graben fault-block rotation initiated during the late Emsian and continued into the Mid Devonian producing the complementary E–W linear Plymouth High (Fig. 10.9b) separating the South Devon Basin from the Looe Basin to the south. This southerly bounding Plymouth High, and the Torquay High that formed a sub-basinal divide to the east, have coeval successions different to those in the basin. However, basin and high successions are linked by clastic sediments derived from the highs, aprons to the volcanic rock edifices associated with the highs extending out into the basins, and sporadic incursions of basinal muds into the successions on the highs. These are detached from their foundations, each being present within two or more northward transported thrust sheets, but generally remain in original relative positions with regard to adjacent basinal successions. The Plymouth High succession is occluded westwards from Plymouth by overthrust deposits of the Looe Basin (Leveridge *et al.* 2002).

Plymouth High succession. The sequence of rocks developed on the high comprises essentially reefal deposits with peripheral carbonate sediments (Plymouth Limestone Formation; Leveridge *et al.* 2002), or, to the east, biogenic bank complex (Brixham Limestone Formation: Drummond 1982; Leveridge *et al.* 2003*b*). The limestones do, however, interdigitate with, or are laterally replaced by, basaltic volcanic rocks (e.g. Ashprington Volcanic Formation).

The Plymouth Limestone Formation is present within two thin thrust sheets, extending east–west through Plymouth. Originally assigned to the Middle Devonian (Ussher 1907) on the basis of a rich coral, brachiopod and gastropod fauna, its rugose corals indicated a Middle–Upper Devonian age to Taylor (1951). Conodonts reveal that it ranges from mid Eifelian to the late Frasnian and possibly the Famennian (Orchard 1977, 1978). A progressive carbonate build-up in the Eifelian produced a blanketed, shallow carbonate platform by the early Givetian, on which coral/stromatoporoid reefs (Orchard 1977) developed with an extensive peripheral facies of reworked crinoids, stromatoporoids and corals, in the late Givetian (Leveridge *et al.* 2002). Succeeding Frasnian deposits, with a distinctive fauna of *Amphipora*, *Stringocephalus* brachiopods and turreted gastropods, desiccation fenestrae and cryptalgal lamination, reflect a restricted and emergent back reef setting (Garland *et al.* 1996) and southward progradation. Fissures, yielding Famennian conodonts (Orchard 1978), indicate foundering of the limestone build-up from mid Frasnian times, termination of carbonate deposition and inundation by red mud deposits.

The Brixham Limestone Formation, present in three thrust sheets (Leveridge *et al.* 2003*a*), is of early/mid Eifelian–Frasnian age. It similarly represents progradation of a biogenic bank complex, but rather than being formed of massive reefs, passes up to massive beds, of *in situ* worked crinoidal debris bound by laminar stromatoporoids, of a reefal flat (Mayall 1979). A Givetian back-reef lagoonal facies, and Frasnian high-energy platform deposits with features of emergence are also present (Drummond 1982).

The Ashprington Volcanic Formation forms an extensive thrust sheet to the south of Totnes in south Devon. Interdigitating with and of similar age to the Brixham Limestone Formation it is a thin slice of a thick sequence of alkaline basalt lavas and tuffs thrust over the South Devon Basin deposits. It represents a major volcanic expression, associated with the Plymouth High, which was emergent, and at times extended into the basin sequences to the north and south (Leveridge *et al.* 2003*a*).

Torquay High succession. The deposits of the high that separated the southern and northern sub-basins of the South Devon Basin, on the eastern side of the province, are present within three or more northerly transported thrust sheets (Leveridge *et al.* 2003*a*). East of the Sticklepath Fault it is represented essentially by the Torquay Limestone Formation, and to the west by the correlative Chercombe Bridge and East Ogwell limestone formations.

The Torquay Limestone Formation, subdivided in part into members, based on lithofacies changes (Scrutton 1977), ranges from early Eifelian (*Polygnathus costatus partitus* Zone: Castle 1982) to early Frasnian (Lower *Polygnathus asymmetricus* Zone: Castle 1982). A similar carbonate build-up to that on the Plymouth High is recorded, with carbonate bank and patch reefs in the Eifelian, giving way to major stromatoporoid reef development in the Givetian, and 'high-energy' conditions in the Frasnian with bioherms within worked crinoidal sand (Scrutton 1977). Locally in the upper part of the formation red silty mudstone, interbedded with and surrounding masses of limestone, yields Mid Frasnian (*Manticoceras cordatum* Zone) ammonoids (House 1963).

The equivalent succession west of the Sticklepath Fault is represented in thrust sheets to the SW of Newton Abbot by the Denbury Crinoidal Limestone, Chercombe Bridge Limestone Formation (ss) and East Ogwell Limestone Formation, and locally developed Kingsteignton Volcanic 'Group' (Selwood *et al.* 1984). Local reef development is characterized by massive corals (e.g. *Disphyllum caespitosum*) rather than stromotoporoids (Scrutton 1977).

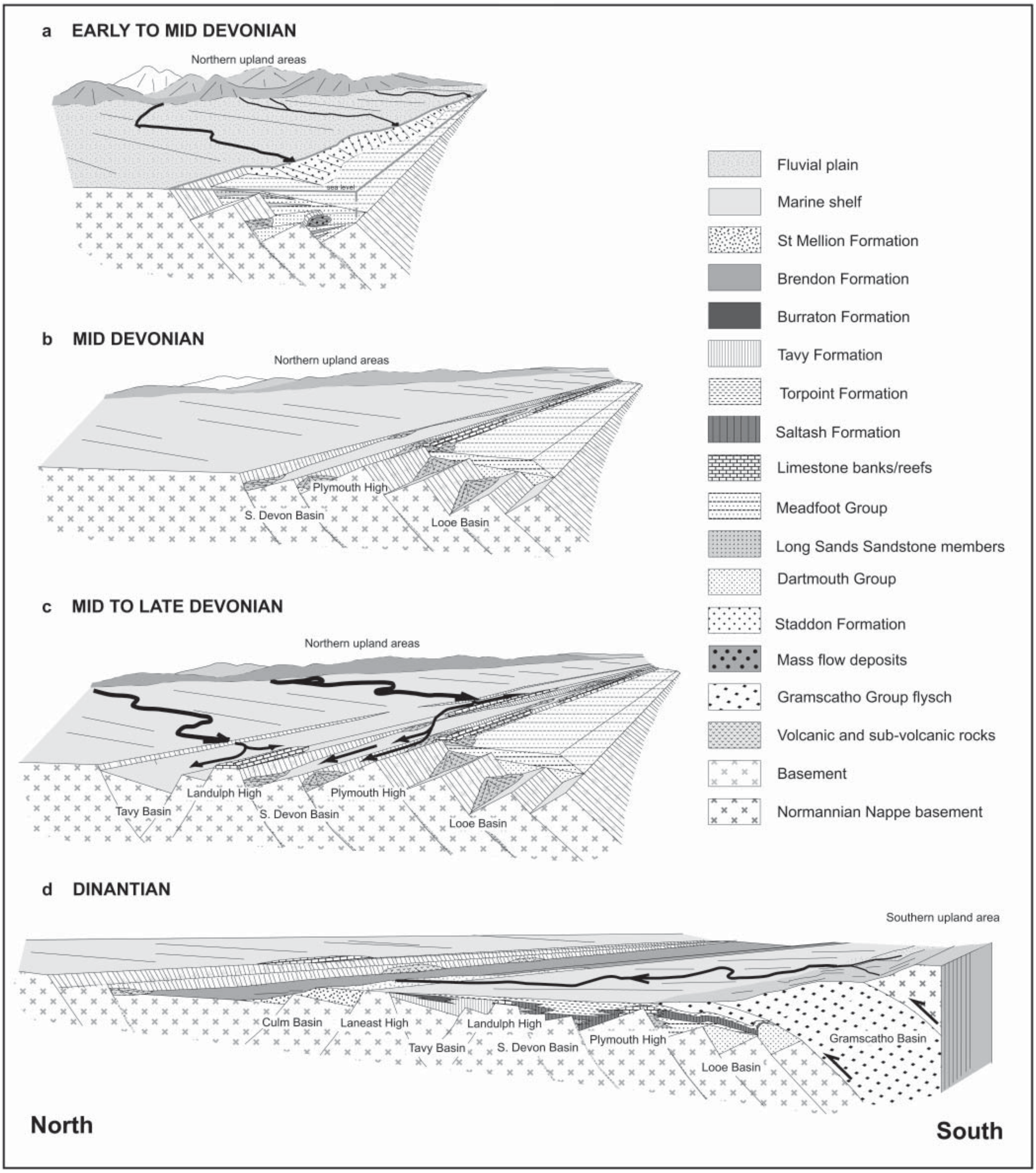


Fig. 10.9. Sequence diagram showing the progressive northward rifting of the passive margin. Sections show basins, sub-basins, highs, derived terrestrial sediment transport and the main lithostratigraphical units of the central part of the sub-province. Based on Leveridge *et al.* (2002).

To the west the Torquay High deposits are occluded by overthrusting, and towards the River Tamar the mid-basin rise was too deep to support biogenic carbonate production (Fig. 10.9b), but related extrusive volcanic rocks and basinal mudstone facies distribution indicates its presence.

Basin succession. The basinal deposits of the South Devon Basin are largely grey, reddish purple and green silty mudstones. Predominant are the grey facies hemipelagic rocks variously termed, from west to east across the province, the Trevose Slate Formation, Saltash Formation and Nordon Formation. Within the South Devon Basin they range from the late Emsian to the Tournaisian in the Saltash Formation of the Plymouth district (Leveridge *et al.* 2002). Upper Devonian occurrences have locally been separately named (e.g. Pentire Slates: Gauss & House 1972) or included with the purple and green mudstones (e.g. Gurrington Slate: Selwood *et al.* 1984). The Trevose Slate Formation (see Selwood *et al.* 1998), some 3.9 km thick, comprises largely submillimetric laminae couplets of fine-grained grey siltstone and dark grey silty mudstone, which is bioturbated and destructured in parts. Subordinate are siltstone and fine-grained sandstone turbidites, and sporadic slump beds. Knox (in Goode & Leveridge 1991) compared the sediments to those of the Santa Barbara Basin in offshore California (Hulsemann & Emery 1961) where laminae couplets are hemipelagic annual deposits in anoxic conditions in a rapidly subsiding barred basin some 500 m deep. Eastwards along the basin laminated sequences become subordinate to homogeneous mudstone. On the southern side of the South Devon Basin, low in the sequence (Eifelian), there are beds of richly fossiliferous siltstone, sandstone and limestone. These are interbedded turbidites in the Plymouth district (Leveridge *et al.* 2002), and storm-generated deposits constituting the separate Bedruthan Formation in North Cornwall (Selwood *et al.* 1998). Crinoidal limestone turbidites form the Marble Cliff Limestone Member (Kirchgasser 1970; Mouravieff 1977), on the western coast near Trevone, an inverted 80 m sequence of Givetian and Early Frasnian age (House *et al.* 1978: Austin *et al.* 1985). Tucker (1969) suggested derivation from a southerly sublittoral environment, possibly the occluded Plymouth High. In the southern sub-basin between Dartington and Torquay (BGS 2003) limestone members of the Nordon Formation are of Eifelian, early Givetian, late Givetian and Famennian age (Drummond 1982). Of similar age to sequence breaks in the Torquay Limestone Formation, they are interpreted as storm current, turbid and debris flow deposits largely derived from the mid-basin Torquay High (Leveridge *et al.* 2003*a*).

Purplish red mudstone and fine-grained siltstone with subordinate green coarse-grained siltstone and fine-grained sandstone constitute the other major sedimentary facies of the basin. Principal occurrences across the basin are the Polzeath Slates to the west, the Torpoint Formation about the Tamar, and the Saltern Cove Formation (and Rora Slate) to the east (Fig. 10.8). Predominantly of Frasnian and Famennian age, there is some evidence of red mudstone being interbedded with late Givetian limestones of the highs, and of it extending into the Tournaisian (Gurrington Slate: Selwood *et al.* 1984). Sedimentary features are indicative of deposition from distal turbid flows and larger suspended turbid clouds (Leveridge *et al.* 2002). Thin mass-flow breccia deposits with Frasnian and Famennian limestone clasts are also locally present at Saltern Cove, Torbay (Van Straaten & Tucker 1972). A sparse benthos but rich planktonic fauna, including entomozoacean ostracodes (e.g. *Nehdentomis serratostriata*) and ammonoids (e.g. *Manticoceras cordatum*) indicate disaerobic bottom conditions in an open marine environment. Water depths of 500–1000 m have been estimated for this same facies, the *Cypridienschiefer*, in southern Rhineland (Bandel & Becker 1975).

The grey hemipelagic facies and the red facies sedimentary rocks are coeval through the Upper Devonian. The two interdigitate on millimetre–kilometre scales, but the red facies peters-out along the basin and sub-basins from areas of maximum development (Leveridge *et al.* 2002). Rich in ferric iron, and a geochemistry indicative of an evolved crustal source, the distribution of red sediments indicate introduction at various points along the northern margin of the basin (Fig. 10.9c). Thickest developments are about the centres of the depositional basin and sub-basins. That they do not form part of a layer-cake succession explains some earlier stratigraphical problems and anomalies of this belt (e.g. Gauss & House 1972; Selwood *et al.* 1993, 1998).

Associated with the basinal sediments there are significant expressions of alkali basaltic lavas, pillowed in parts, hyaloclastites, and tuffs. Major developments are about the mid-basin high (Leveridge *et al.* 2002) and the northern margin of the basin (e.g. Pentire Volcanic Formation: Frasnian; Selwood *et al.* 1998).

Completing the basinal succession about Landulph in the Tamar Valley is a sequence of grey basinal Upper Devonian mudstone passing to dark grey and black mudstone of Tournaisian age, and cherts of Tournaisian and probable Visean age. It is within thin slices derived from the rise slope to the high forming the northern boundary of the South Devon Basin.

Landulph High

Slumped limestone blocks, limestone turbidites, and *in situ* nodular limestones within the sequence at Landulph, indicate the presence of the Landulph High carbonate platform source from the Givetian to the Famennian, subsequently occluded by overthrusting (Leveridge *et al.* 2002). That this was a horst rather than a high generated by block rotation (Fig. 10.9c), is suggested by the associated D_1 structure-confrontation (Leveridge *et al.* 2002). To the east a more complete Late Eifelian–Famennian succession, including the Kingsteignton Volcanic Group, Chercombe Bridge Limestone (s.l.) and the Luxton Nodular Limestone of the Newton Abbot district can be attributed to this high. Part of the Ugbrooke–East Ogwell succession of Selwood *et al.* (1984), it is bound to the SE by overthrust South Devon Basin deposits, and itself is thrust northwestwards over the Kate Brook Slate of the Tavy Basin. Tuffs and lavas of the volcanic group pass up in the early Givetian to limestones that are locally thickly bedded with bioherms up to 6 m thick, and massive stromatoporoidal limestone. The overlying Luxton Nodular Limestone is a condensed mid Frasian and Famennian sequence with a rich pelagic fauna (House & Butcher 1973) representing a marked change of water depth. To the west of the province the high is represented by the Jackets Point Slate Formation (Selwood & Thomas 1986). Comprising mudstone, sandstone, and conglomerate, tuffs and agglomerates, and containing locally abundant shelly faunas, the formation has been interpreted as an association high on the basin shelf margin (Selwood *et al.* 1998).

Tavy Basin

The Tavy Basin (Fig. 10.9c) formed in the Givetian, when the Landulph High became isolated from clastic sediments and developed as a carbonate platform. The principal basinal deposit is green mudstone, the main occurrences across the peninsula being termed the Tredorn Slate Formation (see Selwood *et al.* 1998) in north Cornwall, the Tavy Formation (Leveridge *et al.* 2002) north of Plymouth, and the Kate Brook Slate (Waters 1974) in the Newton Abbot area. The formation comprises pale green–grey-green chloritic silty mudstone, faintly banded in part, with subordinate sandstone, as thin laminated turbidites, sporadic lenses, burrow fills, locally constituting a mappable member. Pyrite is ubiquitous, cubes to 50 mm being common. A restricted benthic macrofauna, including rhynchonelloids and spiriferoids (e.g. *Cyrtospirifer verneuili*) is not definitive, but a late Frasnian–late Famennian age is indicated by its microfossils, conodonts (Stewart 1981*a*), ostracodes (Whiteley 1983) and palynomorphs (Dean 1992). Selwood *et al.* (1998) have proposed an outer shelf setting for the formation, and lithological uniformity of thick sequences to the south led Leveridge *et al.* (2002) to propose rapid sedimentation from turbid flow rather than quiet water settlement from suspension. The chlorite and pyrite content of the rocks indicate the role of reducing conditions during or after deposition, and suggest differing original organic content, burial rates and/or sourcing from the contemporaneous purple facies rocks of the South Devon Basin.

In the northernmost crops of the above dominant facies, there are beds of bioclastic limestone (Selwood *et al.* 1998) and limestone rich sequences, such as the Petherwin and Stourscombe beds (Selwood 1960). These are richly fossilliferous condensed shell concentrate and cephalopod limestones typifying deposition upon or on the flanks of a high, with goniatite, cephalopod and conodont faunas indicating Famennian and early Tournaisian ages (Stewart 1981).

Sporadically intercalated with the predominant facies along the basin are purple and green mudstones (e.g. Rora Slate of the Newton Abbot district) equivalent to those of the South Devon Basin. Planktonic ostracodes show a similar Frasnian (Gooday 1973) to late Famennian (Wilkinson 1990) age range to the major facies. To the west bluish grey mudstone interdigitated with the green constitutes the Delabole Member of the Tredorn Slate Formation (Selwood *et al.* 1998).

Across the basin the green facies rocks pass up to grey facies mudstones such as the Burraton Formation and Yeolmbridge Formation (Selwood 1971). Within these grey–dark grey mudstones, the uppermost deposits of the basin, the transition from the Famennian to the Tournaisian is recorded in goniatite and trilobite faunas (Selwood 1960).

Laneast High

The presence of a high, bounding the Tavy Basin to the north, is intimated, if not fixed, by the limestone sequences of Petherwin and Stourscombe (Stewart 1981*a*, *b*), and possibly by the Buckator Formation (Freshney *et al.* 1972; Selwood *et al.* 1985). Southward-transporting nappes of the Devonian and Carboniferous rocks of the Culm Basin and late extensional faulting in this zone have obscured and occluded much of the body of the high and its deposits. The Laneast Quartzite Formation, of uncertain but probable late Devonian/Dinantian age (Selwood *et al.* 1985), is in thrust contact with Upper Devonian Tavy Basin rocks and basinal Dinantian rocks to the NE of the Bodmin Moor Granite (BGS 1994). Thickly bedded quartzite, coarse-grained, with a high degree of rounding and sorting, is interbedded with subordinate dark grey mudstone with abundant plant debris (Selwood *et al.* 1985). Those authors suggest a high-energy environment of deposition with active reworking, a near shore setting being considered probable. It is suggested here that the formation is a remnant of the shelf bounding the Tavy Basin prior to further rifting and the formation of the Culm Basin that isolated the intervening Laneast High in the Upper Devonian (Fig. 10.9d).

The Petherwin Formation, mid Fammenian-late Tournaisian, Stewart 1981b) and the Stourcombe Formation (late Fammenian, Selwood 1960) are penecontemporaneous deposits on and about the high where it was more deeply submerged. The former formation comprises a variety of lithofacies, nodular cephalopod limestone, calcareous sandstone, fine congomerate, brachiopod-rich, mudstone and the the latter comprises silicified cephalopod (e.g *Gonioclymenia (kalloclymenia)frechi lange, Wochlumeria sphaeroides* (Richter)) limestone, silicified mudstone and trilobite (e.g. *Phacops (phacops)granulatus*(Munster)) rich mudstone. They are characterized by both stratigraphic condensation and reworking in parts. Thus, the earlier shelf sands and contemporaneous deposits of the platform, being variably subject to reworking, sourced sediment into the adjacent basins, and such is manifest in the Buckator Formation (see below: Fig. 10.8)

Culm Basin

The Culm Basin (Figs 10.1 & 10.9) formed as an extensional basin (see below) in the Upper Devonian in sequence with the basins to the south (Leveridge *et al.* 2002). The transition to the Dinantian is recorded within the deposits of both the Tavy and South Devon basins. The main occurrences of Dinantian rocks are present in a polyphase series of E–W thrust slices near the southern margin of the main Carboniferous tract and segregation there of those not attributable to the Culm Basin is presently unfeasible. They are also in thrust nappes, with both Culm Basin and Tavy Basin rocks, that were transported to the south over the Tavy Basin (Fig. 10.1). Sequences originally near the southern margin of the basin are now up to 22 km further south in the St Mellion Outlier (Fig. 10.1) (Whiteley 1983).

Sequences in the thrust sheets extending eastwards from north Cornwall (Freshney *et al.* 1972) north of the Dartmoor Granite (Edmonds *et al.* 1968) to its eastern flank (Selwood *et al.* 1984) are largely similar. Grey mudstone passes up to black mudstone, succeeded by lava and tuff, black mudstone and chert (Fig. 10.8), a sequence generally comparable to the North Devon Basin succession in the paucity of coarse clastic sediments, a result of the 'Bathyal Lull' of Goldring (1962). To the east the greenish grey mudstone of the Trusham Shale yields Tournaisian ostracodes (e.g. *Maternella circumcostata, Richterina (R.) latior*) but the succeeding bluish black mudstone of the Combe Shale has proved barren, attributed to prevailing euxinic conditions by Selwood *et al.* (1984). Equivalent black slaty mudstone at Lydford in the central part of the province contains olistostromes, conglomerates and volcanic debris flows (Isaac *et al.* 1982) indicative of late Tournaisian and Visean instability. On the western coast the black mudstone Barras Nose and Trambley Cove formations has yielded conodonts from limestone lenses indicative of the late Tournaisian–early Visean (Austin & Matthews 1967). Between those two formations is the Tintagel Volcanic Formation, comprising variable proportions of basaltic tuffs and pillow lavas, and represented in the east of the province by lava and high-level microgabbroic sills. There, volcanicity, represented by rhyolitic tuffs and basaltic vesicular lava, extends into the succeeding Teign Chert, and local correlatives. The Teign Chert, consisting of laminae and thin beds of chert, variably coloured, white, black, green and red, and locally rich in radiolaria, is interbedded with black siliceous mudstone. Its Visean age is constrained by a pelagic fauna, in beds of mudstone, chert and limestone at its top, which includes the Zone fossil *Posidonia becheri* (Selwood *et al.* 1984). Similar *Posidonia* beds with the Fire Beacon Chert to the west (McKeown *et al.* 1973) have yielded late Visean bivalves (e.g. *Posidonia corrugata*) and goniatites (e.g. *Mesoglyphioceras granosum*).

Whilst that succession is recognized in occurrences across the peninsula, to the north of Dartmoor and also westwards, the Meldon Shale and Quartzite Formation (Edmonds *et al.* 1968), of probable Tournaisian age (Isaac 1985), consists of grey mudstone with thin beds and lenses of quartzitic sandstone. Such siliciclastic input in the Dinantian is more significant in the Boscastle Formation forming some 4 km of the highly faulted, multithrust, west coastal section between Boscastle and the Rusey Fault Zone (Freshney *et al.* 1972). The formation comprises pyritous black mudstone, with laminated siltstone and packets of greywacke sandstone turbidites. Attributed to the Silesian Crackington Formation by those authors it was reassigned to the Dinantian by Selwood *et al.* (1985), but the early Namurian goniatites (e.g. *Nuculoceras nuculum*) retrieved by Freshney *et al.* (1972) suggest a sequence transitional between the Dinantian and Namurian south of the Rusey Fault.

Interdigitating and stratigraphically conformable with the Boscastle Formation is the Buckator Formation (Freshney *et al.* 1972; Selwood *et al.* 1985), composed of grey-green mudstones with sporadic sandstone beds and limestone lenses with facies-related conodont species indicating a shallow-water sub-tidal depositional setting. The formation ranges from the Fammenian *Scaphignathus velifer* Zone (Selwood *et al.* 1985) to the *Gnathodus bilineatus* Zone (Austin & Matthews 1967).

In the St Mellion Outlier, well to the south of the main crop, three fault-bound Carboniferous formations, the St Mellion Formation, Brendon Formation and Newton Chert Formation are disposed in a series of thrust slices, in an imbricate stack with Torpoint and Tavy formation rocks (Fig. 10.17). In flat lying sheets, transported from the north, bedding is steeply overturned southwards and thick sequences are exposed (Leveridge *et al.* 2002). The major component of the outlier is the St Mellion Formation, composed of sandstone turbidites, slumped plant-rich sandstone, and interbedded dark grey mudstone. Retrieved conodonts indicated a Tournaisian age

(*Siphonella sandbergi* Zone), and a spore flora, including *Lycospora pusilla*, a late Visean age to Whiteley (1983). N. Riley (in Leveridge *et al*. 2002), revising ammonoid identifications by Matthews (1970), determined the presence of the *Nuculoceras nuculum* marine band of the early Namurian. The Brendon Formation, dark-grey siliceous mudstone with sporadic packets of blue-grey coarse-grained greywacke sandstone turbidites, is of largely equivalent age (Tournaisian and Visean: Whiteley 1983) and the Newton Chert Formation, with 'Posidonia Beds' and a rich conodont fauna (Whiteley 1983), compares and correlates with cherts to the north. On sedimentological evidence Jones (1993) described the St Mellion Formation as a pro-delta succession linked to coastal deltaic systems, sourced from the south. The Brendon Formation, juxtaposed by southward thrusting, represents a more distal basinal sequence with sporadic turbidite incursions reaching out into the basin from the delta complex. Abundant plant debris demonstrates a continental source, and grains of low-grade metamorphic rocks and potassium feldspar from exposed granitic rocks, indicate the probable source to have been the up-thrust Normannian and Gramscatho Basin nappes to the south (Fig. 10.9d). The rocks of the St Mellion Outlier are the original marginal facies of the Culm Basin, whereas the Dinantian successions further north represent more distal deposits of the basin (Leveridge *et al*. 2002).

The succession between Boscastle and the Rusey Fault on the west coast, which was also deformed by the main early deformation of the province (see below), is interpreted as the deposits of the southern sub-basin of the Culm Basin (Fig 10.14).

The Dinantian volcanic rocks, Tintagel Volcanic Formation and equivalents, are representative of a wide expression of volcanism in the Rhenohercynian Zone. It has been linked to high silica content in basinal waters promoting formation of associated radiolarian chert (Ziegler 1982). As with most of the Devonian volcanic rocks they are alkaline basalts (Floyd 1984) with Ocean Island Basalt characteristics and a mantle source (Merriman *et al*. 2000), typical of an extending rift regime with β (stretch factor) > 2 (McKenzie & Bickle 1988). The Culm Basin thus originated as one of the passive margin rift basins. The volcanics are also the last of the province, and of the extensional regime that characterized most of the Devonian and Dinantian prior to regional shortening, basin and conduit closures and deformation.

North Devon 'Basin'

The Devonian–Dinantian rocks of north Devon and west Somerset (Fig. 10.10) constitute the succession of the North Devon Basin (Fig. 10.1). The limits and nature of this depositional basin are very poorly constrained. The area has been considered (e.g. Edmonds *et al*. 1975; Tunbridge 1978; Selwood & Durrance 1982) to have lain immediately to the south of Wales with the transition from the continental Old Red Sandstone facies to the Devonian marine facies being accommodated in the Bristol Channel area. Certainly, through the Devonian, its rocks do reflect the interplay of those facies. However, it has been proposed that such interplay took place up to 400 km to the SE (Holder & Leveridge 1986a, 1987). Sequences would therefore equate with the 'Old Red Sandstone' (Eastern Avalonia) derived rocks of the southern Ardennes (Behr *et al*. 1984), before Carboniferous displacement along the Bristol Channel–Bray Fault (Kellaway & Hancock 1983; Higgs 1986, fig. 2). The succession, although deformed by major–minor E–W northward-verging and upright gently plunging folds (Sanderson & Dearman 1973) with an axial plane slaty cleavage, youngs from north to south in a simple manner (Fig. 10.10). Edmonds *et al*. (1985) disputed the contention of others (e.g. Reading 1965; Holwill *et al*. 1969) that thrusting repeated sequences significantly, and suggested a minimum total thickness for the succession of approximately 6700 m (Fig. 10.11). Limited evidence from the Devonian of south Devon points to thin shelf expressions to the north of lithostratigraphical divisions formed mainly in the developing

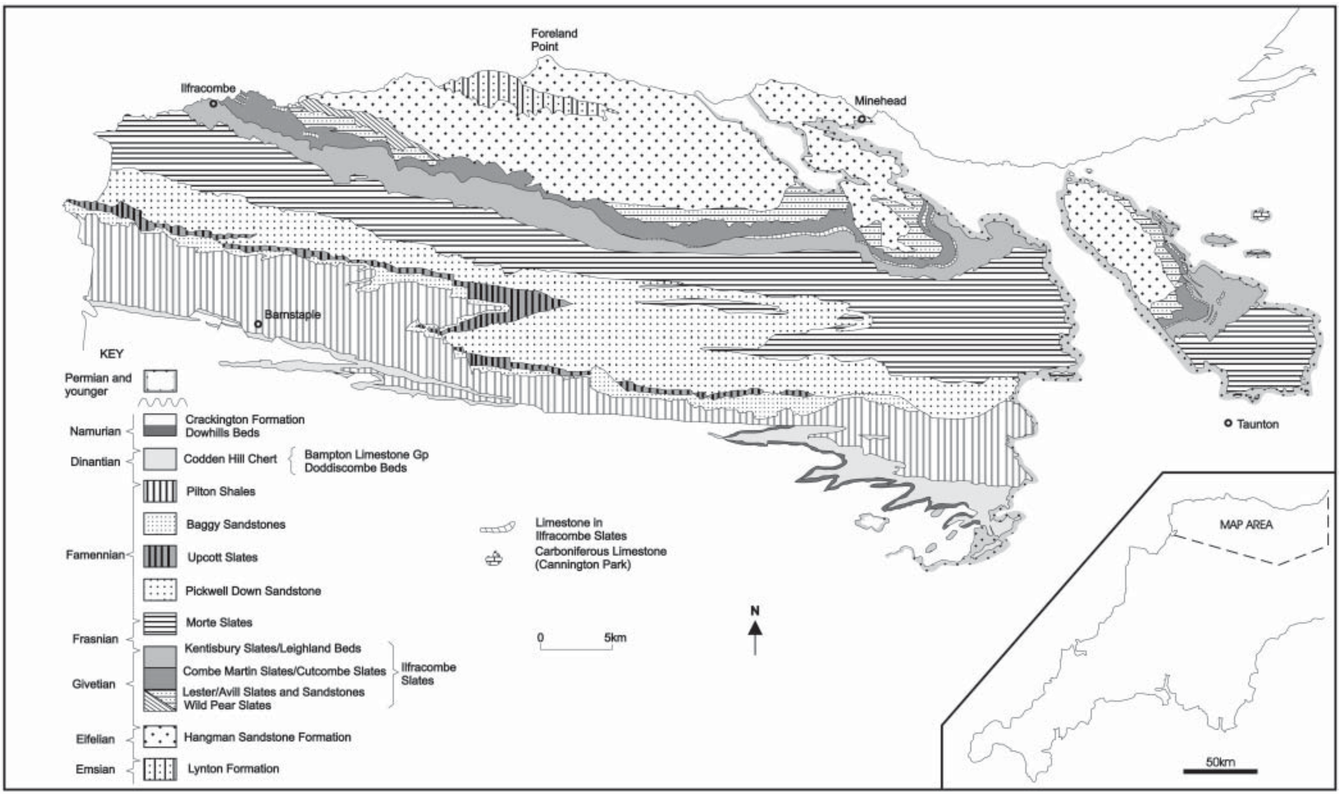


Fig. 10.10. Geological sketch map of the north crop Devonian and Dinantian succession, based on BGS maps.

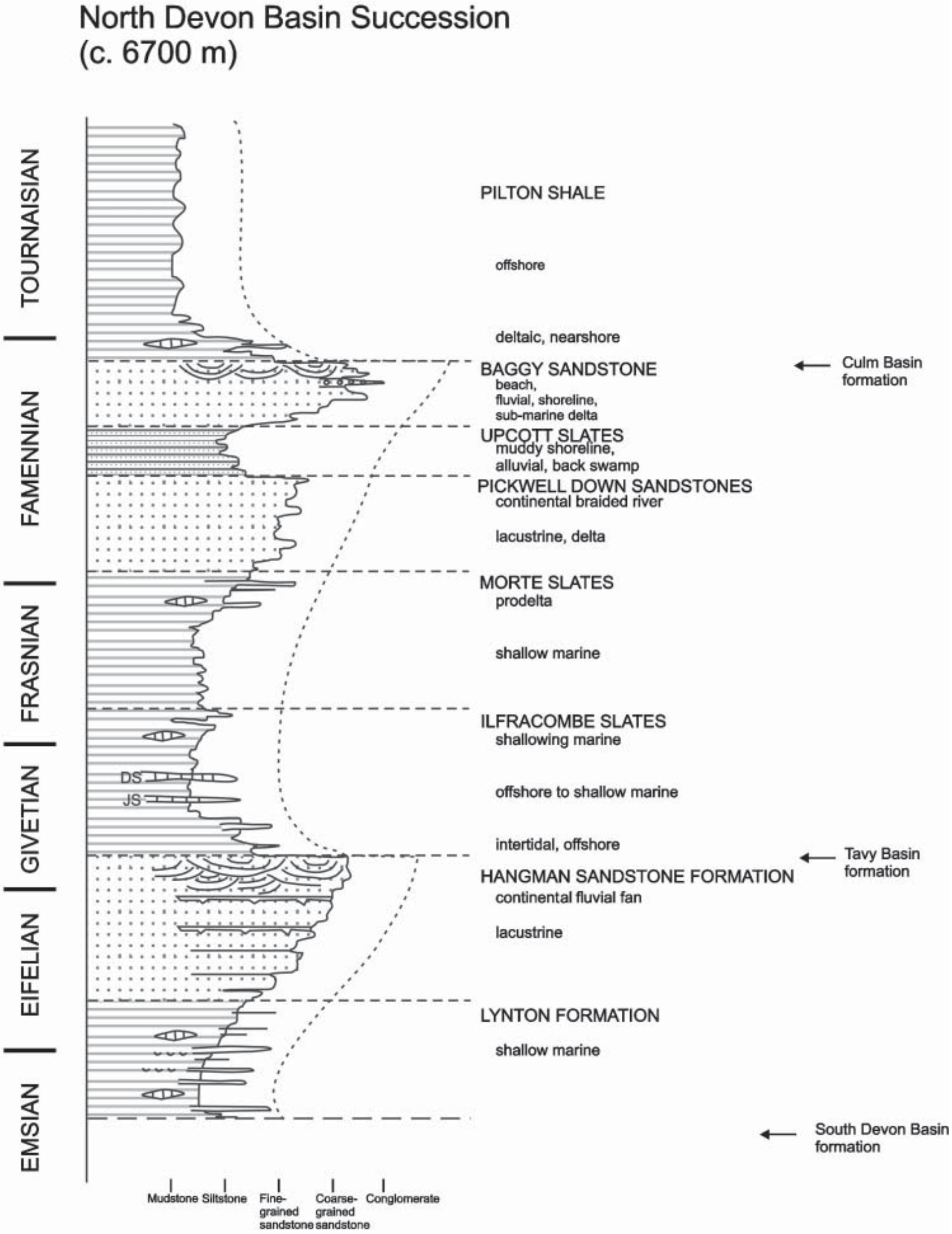


Fig. 10.11. Graphic log of the North Devon Basin succession.

basins (Leveridge *et al.* 2003a). This suggests that the main depositional axis of this basin was located in the north where sediment debouched before an elevated northerly hinterland.

The oldest rocks of the north crop form the Lynton Formation (Edwards 1999), lying within the faulted core of the Lynton Anticline, a major fold plunging gently east-southeastwards from Lynmouth Bay. The formation comprises largely interlaminated grey mudstone and sandstone, with the latter predominating in the lower and uppermost parts, also included are beds of quartzite and thin beds and lenses of bioclastic limestone. Bioturbation and burrows, including *Chondrites* (Simpson 1964) are common, and bivalves, brachiopods, bryozoans and crinoid debris constitute the limestone. Brachiopods including *Platyorthis longisulcata* and *Chonetes sarcinulatus* indicated a late Emsian–early Eifelian age to Evans (1983), a range confirmed by the shallow-water conodont faunas retrieved by Knight (1990). The shallow-marine setting of the deposits (Simpson 1964; Edmonds *et al.* 1985) was considered by Selwood & Durrance (1982) to represent 'almost certainly' the initial marine transgression of the Old Red Sandstone continent.

Re-establishment of continental conditions is represented in the succeeding Hangman Sandstone Formation (Edwards 1999). Variously estimated to be between approximately 1650 (Tunbridge 1978) and 2500 m (Edmonds *et al.* 1985) thick, the formation comprises predominantly purple, green and grey fine- to coarse-grained sandstone, with subordinate reddish brown mudstone. The mudstone, variably finely interlaminated with sandstone or homogeneous with extrabasinal clasts or pedogenic carbonate nodules (Jones 1995), shows desiccation cracking and bioturbation. Plant stem fragments, trace fossils, *Arenicolites* and the non-marine *Beaconites*, and gastropods *Naticopsis* and *Bellerophon* occur sporadically through the formation, and the bivalve *Myalina* is common towards the top, but none are definitive of age. However, palynomorph assemblages from the upper part of the formation have indicated a late Eifelian/early Givetian age to Knight (1990). Sedimentological analyses by Tunbridge (1981*a*, 1984) and Jones (1995) have defined a coarser facies of channel and sheet sandstones, with palaeocurrents indicating transport to the south, and a finer facies of mudstones deposited in lakes, or as terrestrial mass flows. Deposition was largely within a distal fluvial fan environment (Jones 1995) into which there were brief marine incursions and on which lakes and mudflats developed at intervals (Edwards 1999). Heterolithic beds and cross-bedded sandstone, with bimodal current directions, and the presence of locally abundant *Myalina* and bryozoans in the uppermost part of the formation, herald the next major marine incursion of the area (Tunbridge 1978).

The establishment of fully marine conditions is recorded in the Ilfracombe Slates. Mappable subdivisions (Edmonds *et al.* 1985) of the Slates (Fig. 10.10), similar to those described by Holwill (1963), reflect the deepening and shallowing of that marine environment. The silver-grey mudstone, with thin beds of sandstone siltstone and limestone of the Wild Pear Slates indicate shallow-water deposition (Edmonds *et al.* 1985), with a W–E along-strike variation from delta front to intertidal suggested by Webby (1965*a*). The Lester Slates-and-Sandstones contain a greater proportion of cross-bedded sandstone and thin bioclastic limestone beds within the mudstone, in which *Chondrites* is abundant. The sandstones and shelly limestones, with coral, brachiopod, bryozoan and crinoid debris, are interpreted as storm generated deposits in an offshore setting. The Combe Martin Slates comprise the silver-grey mudstone with sporadic thin sandstone beds and limestone beds, the latter forming two units up to 10 m thick mapped across the province (Fig. 10.10), namely the lower Jenny Start Limestone and the higher David's Stone Limestone. Corals and stromatoporoids, commonly in growth positions, characterize both. Rugose corals, such as *Disphyllum aequiseptatum* and *Thamnophyllum caespitosum*, are preponderant over tabulate corals in the lower unit, and the reverse relationship applies in the upper unit, with tabulate corals such as *Thamnopera cervicornis*, exceeding single corals. This has been regarded as an indicator of shallowing in the sequence (Webby 1966; Edmonds *et al.* 1985). The Givetian age indicated by the corals is confirmed by conodonts assigned to the Lower/Middle *Polygnathus varcus* Subzone (Knight 1990). The superceding Kentisbury Slates, comprising silver-grey mudstone and subordinate sandstone rich units, the former characterized by *Chondrites* burrows, are interpreted as representing regression in the early Frasnian (Edmonds *et al.* 1985). Within the division there are argentiferous lead deposits extending across the crop from Combe Martin, regarded by Scrivener & Bennett (1980) as sedimentary exhalative (Sedex) deposits linked to the Cockercombe Tuff (Webby 1965*b*) in the Quantock Hills.

How the succeeding Morte Slates relate to the cycles of regression and transgression is uncertain. Comprising some 1500 m of silver, grey-green and purple mudstone, sporadic thin cross-laminated sandstone beds in lowermost and central parts, and sporadic limestone nodules, they carry a sparse impoverished fauna, which includes *Lingula* and *Cyrtospirifer verneuili*. They are regarded as shallow-water marine deposits, with a continued regression (Selwood & Durrance 1982), or further transgression (Edmonds *et al.* 1985), represented by the prodelta and delta platform setting proposed by Webby (1965*c*, 1966). The depleted macrofauna, suggested by Selwood & Durrance (1982) to be linked to rapid sedimentation and reduced salinity at a delta mouth, is reflected in the depletion of the miospore and acritarch assemblages, that typify the Frasnian and earliest Famennian (Knight 1990).

The overlying Pickwell Down Sandstones consist of purple, red, brown and greenish grey sandstone, with subordinate red and grey mudstone near the top and base of the formation. Also, at the base is the Bittadon 'Felsite', a tuff up to 8 m thick that extends across the north crop, that has yielded remains of Upper Devonian armoured fish such as *Bothriolepis, Holoptychius and Coccosteus* (Rogers 1919). The main body of the 1200 m-thick formation comprises fine- to medium-grained thick to massive sandstone beds with wavy and cross-lamination, trough cross-bedding, erosive bases, and ripple marking. It is considered (Selwood & Durrance 1982; Edmonds *et al.* 1985) to represent predominantly continental braided river deposits, with subordinate lacustrine and deltaic facies. The formation contains few macrofossils apart from sporadic fish and plant remains, but it is assigned to the Famennian as it is in sequence with dated mudstones below and higher in the succession. There is upward passage to the Upcott Slates, a sequence of mudstones and silty mudstones, variegated cream, buff, green, grey and purple, with sporadic thin fine-grained cross-laminated sandstone beds, that is up to 250 m thick. The slates were thought by Goldring (1971) to have been deposited in alluvial, back-swamp, lacustrine, and muddy shoreline environments.

The Baggy Sandstones, overlying the Upcott Slates, comprise up to 450 m of sandstone, thick cross-bedded cosets to thinly bedded, siltstone and mudstone, with minor developments of intraformational conglomerate and slumped sediments, and sporadic thin beds of crinoidal and gastropodal limestone. Relatively rich in macrofossils and trace fossils, Goldring (1971) considered it an exemplar of shallow-water sedimentation. Within the framework of preponderant fine-grained facies associations towards top and base, and sandstone facies dominating in the main part of the formation, he defined a variety of subfacies and linked depositional environments. Prevailing submarine delta platform sedimentation is interrupted by repeated shoreline advances, represented by a diversity of deposits in fluvial distributary, estuarine channel, lagoonal and beach settings. A close relationship between environments and fossils was defined: plant debris (*Sphenopteridium rigidum* and *Knorria*) with distributaries and lakes, trace fossils (*Arenicolites curvata*, *Diplocraterion yoyo*) with a sub-beach setting, gastropods and bivalves (e.g. *Dolabra 'Cucullaea' unilateralis*) with near shore intertidal channels, and productellids and bivalves (e.g. *Ptychopteria damnoniensis*) with open marine conditions. The upper boundary of the Baggy Sandstones with the Pilton Shales is transitional and poorly defined (Edmonds *et al.* 1985). However, Goldring (1971) reported palynomorph and conodont evidence that the formation extended to the topmost Devonian '*Wocklumeria*' Stufe. Conodont analysis by Austin *et al.* (1970) indicated that a younger late Devonian biozone (Lower *Bispathodus costatus* Zone: Austin *et al.* 1985) extended across the boundary of the formations.

The lithostratigraphical base of the Pilton Shales is arbitrarily taken to be immediately above the uppermost sandstone unit attributable to the Baggy Sandstones (Edmonds *et al.* 1985). The formation extends up to the dark grey and black shaly mudstone of the succeeding Codden Hill Chert and correlatives within the Dinantian. Some 500 m thick, the formation comprises grey shaly mudstone, with thin beds of siltstone and sandstone, thin beds and lenses of limestone, and sporadic calcareous sandstone beds and composite units in its lower part. The sequence is rich in fossils (see Goldring 1970; Edmonds *et al.* 1985), particularly brachiopods, bivalves, trilobites and goniatites. Goldring (1970) linked lithological and faunal changes through the formation to major transgression, with residual near shore and deltaic sediments and associated brachiopods and bivalves, giving way to an offshore finer facies with trilobites and goniatites. The lower third of the formation Goldring (1957, 1970) attributed to the Famennian *Wocklumeria* Stufe on the basis of its productellid and trilobite (e.g. *Phacops accipitrinus accipitrinus*) fauna. The overlying sequence represented the Dinantian *Gattendorfia* and possiby the *Ammonellipsites* ammonoid zones (Tournaisian–early Visean).

The Devonian succession (Fig. 10.11) constitutes two major sedimentary cycles, with the timing of marine transgressions coinciding with basin formation to the south. This suggests general subsidence of the passive margin accompanying the release of extensional stress.

The Dinantian succession of the north crop is completed by a sequence of dark grey and black shaly mudstone, chert and limestone. These are designated the Codden Hill Chert (to the west and north of the E–W Brushford Fault), and the Dowhills Beds and Bampton Limestone 'Group' (to the east and south of the Brushford Fault). The Codden Hill Chert comprises some 150 m of dark grey–black shaly mudstone, locally siliceous, and chert, each of which is locally predominant (Prentice 1960*a*), with sporadic lenses of limestone. The chert, in part rich in radiolaria, is generally pale weathering. *Posidonia becheri* is common through the sequence, and both trilobites and goniatites (e.g. *Goniatites spiralis*) confirm a Mid–Late Visean attribution (P_1–P_2: Prentice 1960*a*). Correlatives to the east are the Doddiscombe Beds and Bampton Limestone Group (including the Westleigh Limestones). The Doddiscombe Beds, comprise pale and dark grey laminated mudstone, which is unfossiliferous and variably cleaved. The underlying Pilton Shales yield Tournaisian *Siphonodella* Zone conodonts (Matthews & Thomas 1968) and the superceding Bampton Limestone Group is of Visean age (Thomas 1963*a*). The Bampton sequence consists of interbedded chert, limestone and shale. Further to the east at Westleigh turbidite limestone beds, up to 6 m thick and locally conglomeratic, are interbedded with the shaly laminated mudstone. Pelagic (orthocones and *Posidonia*) and subordinate benthonic faunas in the shales, indicative of accumulation in quiet water conditions, contrast with the oolitic material, rolled coral colonies and abraded brachiopods in the limestone beds derived from a shallow higher energy shelf regime (Thomas 1963*b*).

Continental collision and deformation of the passive margin

The successions of the southern Devonian–Dinantian sub-province are complexly deformed in a fold and thrust terrain (e.g. Isaac *et al.* 1982; Coward & McClay 1983). Early fold facing and thrust transport are northwards in a southern domain and southwards in a northern domain (e.g. Sanderson & Dearman 1973; Seago & Chapman 1988). This overall structure of the belt was interpreted in terms of forward propagating thrusts originating in the south, with backfolding and backthrusting of an essentially layer-cake stratigraphy by Coward & Smallwood (1984) and Seago & Chapman (1988). The significant role of basin architecture in influencing structure development during basin inversion was proposed by Selwood (1990), who attributed southward-verging structures in the Trevone Basin (South Devon Basin) to backfolding associated with basin inversion and northward out-thrusting on its southerly inclined northern margin fault. Hartley & Warr (1990) and Warr (1993) also attributed southerly backthrusting, and consequently the facing confrontation of the Trevone Basin (e.g. Roberts & Sanderson 1971; Selwood *et al.* 1993), to the butressing effect of a rise bounding the rift basin to the north. The controlling influence of basin architecture on the development of major structures during inversion, with basin margin faults, not only becoming thrusts bounding basinal successions, but also controlling regional structure facing and structure confrontations, was proposed by Leveridge *et al.* (2002).

Closure of the Gramscatho Basin was accommodated by northward tectonic migration (see above; Fig. 10.7). The concept that such migration operated through the whole of the southern crop of Devonian and Dinantian rocks (e.g. Dearman 1971; Shackleton *et al.* 1982) was founded upon a perceived structural continuity, and a general progressive northward younging of metamorphic radiometric ages. Structural continuity is now established through this sub-province, structural sequences in both active and passive margin successions being similar, with three major deformation episodes (e.g. Alexander & Shail 1996; Leveridge *et al.* 2002) common to both, and subordinate phases locally developed (Leveridge *et al.* 1990; Selwood *et al.* 1998). The whole-rock K–Ar dating of Dodson & Rex (1971), adjusted by Warr *et al.* (1991) using revised decay constants, remains an important constraint on the setting-time of metamorphism. The low-greenschist grade of metamorphism within the belt is considered to be a function of both burial and thrust nappe loading (Warr *et al.* 1991), with slaty cleavage development of the first deformation, D_1, promoting strain enhanced crystal growth for high anchizonal and epizonal grades (cf. Merriman *et al.* 1995). The age range of slates within the belt is 345–325 Ma, spanning the late Tournaisian–early Namurian, and this is when the passive margin was first deformed. Younger narrow E–W zones along the northern margin of the belt (295–280 Ma) and between the Start peninsula and Perranporth area (315–295 Ma) are attributable to secondary (D_2) transpositions of earlier fabrics (e.g. Holdsworth 1989; Pamplin 1990).

Collision to the south in late Famennian/early Dinantian times was thus not a terminal event, and deformation migrated northwards. Structures of D_1 in the passive margin basin terrain are an ubiquitous penetrative cleavage, close to isoclinal folds, thrusts and strike-slip faults on mesoscopic and macroscopic scales. The prevailing orientation of folds and thrusts is E–W (e.g. Alexander & Shail 1996; Selwood *et al.* 1998), but there is incipient sheath folding, particularly in the vicinity of major thrusts and in higher strain zones to north and south (Dearman 1969; Leveridge *et al.* 1990). Related extension lineations are orientated NNW–SSE. Major D_1 thrusts bound basin and sub-basin successions (Leveridge *et al.* 2002). They have major antiformal folds in hanging walls at thrust fronts, a notable example being the Bovisand/Man Sands Antiform at the northern margin of the Looe Basin succession. This is a product of the inversion of half-graben and full-graben and the translation of each bounding rift fault into a basal thrust (Fig. 10.12). The process progressed with each extensional basin closing, inverting, and 'locking-up' before stress transmission to its neighbour

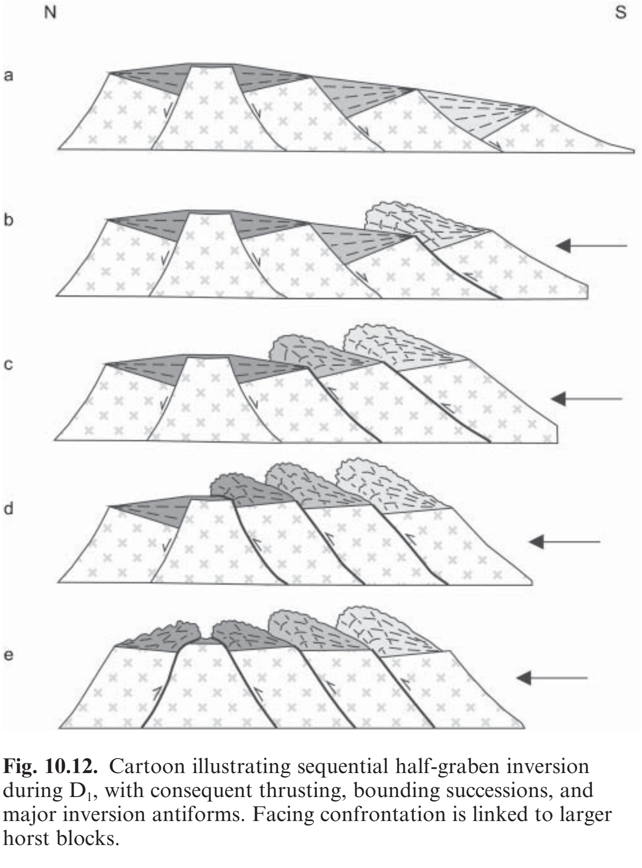


Fig. 10.12. Cartoon illustrating sequential half-graben inversion during D_1, with consequent thrusting, bounding successions, and major inversion antiforms. Facing confrontation is linked to larger horst blocks.

(Leveridge *et al.* 2002). Basin contents were not expelled over the major highs in south Devon, but the successions of the highs (Plymouth and Torquay) were detached, apparently by footwall short-cut thrusts, and translated over adjacent deposits to the north in the South Devon Basin (Fig. 10.15) (Leveridge *et al.* 2003*a*). In a similar manner the reversal of facing at or near the southern margin of the Tavy Basin (Selwood *et al.* 1984; Leveridge *et al.* 2002) within the northern domain is attributable to a northerly inclined bounding fault at the Landulph High. D_1 structures face southwards over much of the crop of the Tavy Basin deposits (Seago & Chapman 1988; Selwood *et al.* 1998). A further reversal of facing in the north (Isaac *et al.* 1982), where largely obscured by overthrust Culm Basin deposits, probably reflects a full-graben geometry for the basin. The facing confrontation between South Devon Basin and Tavy Basin successions across much of the peninsula is accommodated without interference, D_1 cleavage being common across the zone (Leveridge *et al.* 2002).

The confrontation of early structure facing on the north Cornwall coast, variously known as the Polzeath or Padstow Facing Confrontation, is within the South Devon Basin (Trevone Basin; see Selwood *et al.* 1998). Subsequent to its description by Gauss (1967) as two separate tectonic domains juxtaposed by later thrusting, and as a zone of overlap of northward and southward facing structures (Roberts & Sanderson 1971), debate centred on the relative timing of those early structures (e.g. Pamplin & Andrews 1988; Durning 1989). The confrontation has been attributed to backfolding (Andrews *et al.* 1988) and to backfolding and thrusting associated with basin inversion (Selwood 1990) or backthrusting caused by buttressing against earlier extensional faults to the north of the basin (Warr 1993), all mechanisms being ill constrained. Overlap fabrics of the zone can now be largely attributed to regional D_2 deformation associated with the northward transporting Trebetherick Thrust, that juxtaposes reversed facing, but beneath which the reversal-transition of primary facing is recorded (Selwood *et al.* 1998). The coeval, but locally distinct facies to north and south in the basin (Gauss & House 1972; Selwood *et al.* 1993) reflect division of the basin apparent to the east. The confrontation is compatible with inversion about a horst-block sub-basinal divide.

During deformation of the Devonian rocks there was deposition within the Culm Basin, with coarse clastics at the southern margin of its southern half-graben, passing northwards to finer grained condensed basinal deposits. At least the southern part of the basinal succession, up to and including early Namurian sediments, was then deformed in sequence by D_1, while the spread of thrust sheets of this succession across the basins to the south is in large part a result of D_2 and D_3 (see below). However, the southward facing and overturning of all these rocks (Seago & Chapman 1988; Leveridge *et al.* 2002; BGS unpublished data), rather than northward (Isaac *et al.* 1982), is a D_1 feature, resulting from the formation of major southward-verging antiformal folding at the southern basin margin during inversion. D_1 structures, overturned and southward facing, are developed (with D_2) within the Boscastle Formation (Selwood *et al.* 1985) up to the Rusey Fault zone (Freshney *et al.* 1972). The opposed vergence of structures in Culm Basin rocks, southwards on the southern margin, and northwards to the north (Dearman 1971), although including Silesian rocks deformed after the D_1 inversion to the south, is an indicator that the extensional basin was a full graben.

The D_1 transport direction, between north and NNW, is approximately normal to the main basinal architectural elements of the province, and thus transpressive fabrics are not typical of the deformation. At a late stage of D_1 the prominent NW–SE dextral strike-slip faults of the province developed (Leveridge *et al.* 2002), possibly as the extensional basins became largely locked. They are part of a wide family of stuctures across the Variscan belt, which includes the Bristol Channel–Bray Fault (Fig. 10.2), and along which associated ductile tectonic reworking has been dated at approximately 320 Ma (Matte *et al.* 1986), mid Namurian.

The Silesian Culm Basin

The major proportion of the Namurian and all of the Westphalian rocks of the Culm Basin had not been deposited when the Dinantian and lowermost Namurian succession near its southern margin was inverted by D_1. Continuing tectonic migration, consequent uplift and sedimentation were therefore concurrent (Warr 1993). With progressive closure of the extensional Culm Basin and the presence of inverted basins to the south, the Culm depositional area became a foreland basin during the Silesian (Hartley & Warr 1990; Hartley 1993*b*).

The principal formations of the foreland Culm Basin are the Crackington, Bude and Bideford formations (Edmonds 1974; Thomas 1988). Early Namurian mudstone sequences, such as the Dowhills Beds (Webby & Thomas 1965) to the north, Ashton Formation (Chesher 1968) to the south, are essentially conformable with Dinantian sequences and products of the bathyal lull, prior to the incursion of coarser synorogenic clastics of the main crop (Fig. 10.13). They have been regarded as sandstone-free, or sandstone-sparse, parts of the Crackington Formation (Edmonds 1974). The Bealsmill Formation between Dartmoor and Bodmin Moor (BGS 1994), constituting the Blackdown Nappe (Isaac 1985), derived from the north (Seago & Chapman 1988; BGS unpublished data), is a proximal facies (Selwood & Thomas 1988) correlative of the Crackington Formation. The affinity of the Early–Mid Namurian (House & Butcher 1973) proximal submarine-fan deposits of the Ugbrooke Sandstone in the Newton Abbot district, to either the Crackington Formation or the St Mellion Formation, is uncertain.

The Crackington Formation, lowermost of the formations of the Culm Basin, is present at the southern side and northern side of, and in antiformal inliers between, the main crop of the Silesian (Fig. 10.13). It comprises interbedded sandstone and dark grey–black pyritous shaly mudstone, in locally variable proportions, with sandstone becoming dominant higher in the sequence. The sandstone is a fine-grained quartz-rich greywacke forming laminae to thick beds, predominantly medium beds, of increasing thickness upwards (McKeown *et al.* 1973). Interpreted as distal turbidites, most beds are massive with delayed grading, others show distribution grading, and commonly beds have sharply defined bases with load casting and cross-laminated tops. Sole marks, groove casts and flute molds, prod marks, largely indicate easterly or westerly flow, with subordinate transport directions to the NE and south (Mackintosh 1964; Edmonds *et al.* 1968). The formation is characterized by a restricted marine benthos, ammonoids, fish and bivalves, and dispersed plant debris. Turbidite deposition commenced in the late Early Namurian (*Eumorphoceras* (E_2) subzone: Edmonds *et al.* 1968) in the south and generally in the early Late Namurian (*Reticuloceras* (R_2) subzone: Webby & Thomas 1965) in the north. Exceptionally, at Venn in the north (Edmonds *et al.* 1985), one sandstone occurrence yields *Homoceras* (Mid Namurian). It continued into the Early Westphalian ('A'; *Gastrioceras amaliae* subzone: Edmonds *et al.* 1979). Sedimentation is interpreted to have been in 'relatively' deep water, largely anoxic, with turbidite flow predominantly parallel to the basin axis. Sourcing has been attributed in small part to collapse from the inverted southern basin margin (see Thomas 1988), and largely to an emergent continental terrain to the north (Melvin 1986) and east (Freshney & Taylor 1980), possibly linked to the contemporaneous activity of the Bristol Channel/Bray Fault.

The Bideford Formation has a relatively restricted east–west linear distribution extending from the coast just to the south of Westward Ho! It is composed of coarsening-up sedimentary cycles (De Raaf *et al.* 1965), from black mudstone with sporadic

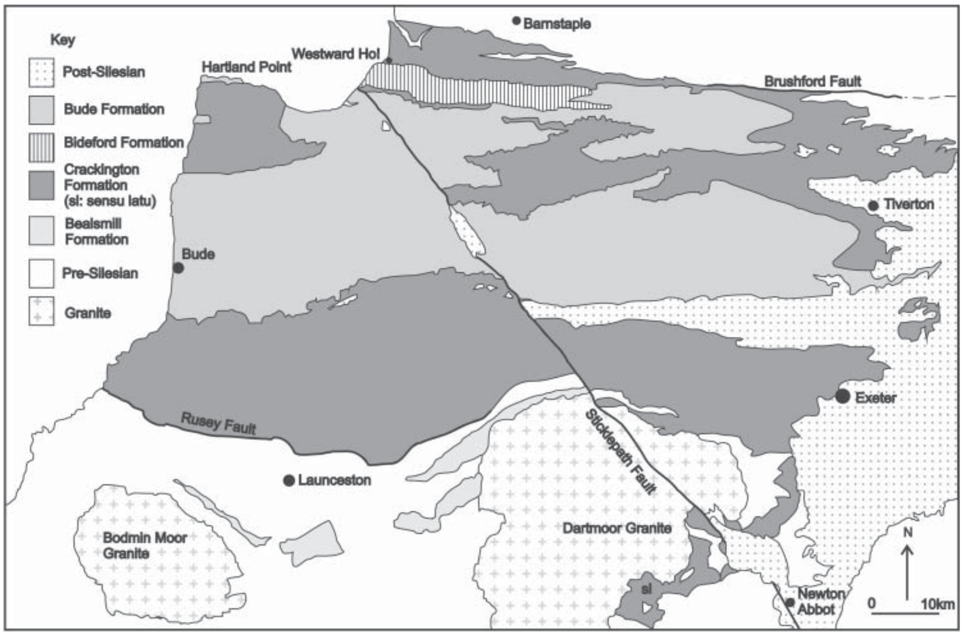


Fig. 10.13. Sketch geological map showing the distribution of the lithostratigraphical divisions of the Silesian, and the main mapped faults. Based on BGS maps, and Thomas (1988).

distal sandstone turbidites, through silty mudstone, siltstone, siltstone with sandstone, to medium-grained feldspathic sandstone. The upper sandstone units, sourced from the north (Prentice 1962) are commonly cross-bedded with channel features or wave-rippling, and rarely capped by rootlet beds, seat earths and thin seams of coal. A non-marine fauna (e.g. *Carbonicola*) is present at intervals through the sequence, but scarce marine fossils are present. *Gastrioceras amaliae* is recorded high in the formation (Edmonds *et al.* 1979) indicating its correlation in part with the Crackington Formation and the Bude Formation. The sequence represents sedimentation in a deltaic setting (Prentice 1960b), with associated nearshore and shoreline beach facies (Elliott 1976), formed in a basin sheltered from vigorous wave action.

The Bude Formation succeeds the Crackington Formation conformably over much of the basin. It is characterized by thick beds (to 5 m) and amalgamated units (to 20 m) of poorly sorted feldspathic sandstone with extrabasinal clasts (Freshney *et al.* 1979*a*, *b*), dark laminated mudstone and siltstone with interbeds of graded sandstone, black shaly mudstone, and 'slumped beds' (Freshney *et al.* 1972). Amalgamated units were considered by Melvin (1986) to lack organization, but Higgs (1991) identified coarsening up/fining up sequences as typical. Sandstone beds are locally erosive, bases exhibiting grove casts and flute molds, indicative of palaeocurrent transport from the north. Sporadically developed are sandstone beds with abundant plant remains and coal laminae (Edmonds *et al.* 1968), and rain spotting is recorded on some surfaces (Lloyd & Chinnery 2002). King (1966) interpreted locally abundant trace fossils as xiphosurid (king crab) trails, an indicator of a shallow-water setting. Fauna is generally sparse, apart from within a few extensive black shale 'marine bands' (Freshney & Taylor 1972) rich in goniatites. These indicate that the formation extends above the *Anthracoceras aegiranum* horizon and the Westphalian 'B/C' junction (Thomas 1988). With background sediments, including distal turbidites, indicating some continuity of the Crackington conditions, interpretations of the sedimentary environment of the formation based on sandstone sedimentology have varied between fluviodeltaic (e.g. Prentice 1962) and a deep-water (e.g. Lovell 1965) setting. A more recent consensus, that sandstone deposition was from turbidity currents, was not extended to the setting. A shallow basin pro-delta slope environment was favoured by Melvin (1986), supplied by a delta to the north, with levéed channels, and marine bands representing sedimentation after lobe abandonment. Higgs (1991) proposed a major freshwater lake setting, in which fluvial-fed sandstone turbidites were deposited above storm wave-base, and marine incursions occurring only at times of high eustatic sea level. However, the coaly deposits recorded by Edmonds *et al.* (1968) in the southern synformal crop of the Bude Formation, east of the Sticklepath Fault, may also indicate deltaic progradation across part of the basin late in the development of the formation.

The configuration of the extensional Culm Basin as a broadly symmetrical full-graben, with a deeper inner basin, flanked by shallower half-graben sub-basins (Fig. 10.14), is constrained by its inversion structures, and the facies distribution of its infilling sediments (see below). The early–mid Namurian inversion of the southern sub-basin left the inner-basin, bound to the south by the Rusey Fault, and northern sub-basin as the main areas of Silesian deposition. Continuing D_1 contraction resulted in progressive inversion of these basins during the Namurian and Westphalian (Fig. 10.14), with sedimentation in increasingly shallow water. Northerly-directed synsedimentary thrusting, with some décollments in units previously termed 'slump beds' in the Bude Formation, has been detailed (Enfield *et al.* 1985; Whalley & Lloyd 1986; Lloyd & Chinnery 2002). A very localized incipient cleavage appears to relate this deformation to ongoing D_1 in the regional structure sequence. The southerly inverted area effectively shielded the main basinal area from southerly sources, the Bealsmill Formation probably representing trapped sediments that were subsequently out-thrust to the south. Emergence of a source to the north supplied sediment to the basin in the Namurian (Hartley & Warr 1990). This is compatible with northerly overthrusting along the E–W Bristol Channel section (Brooks *et al.* 1993; Edwards 1999) of the Bristol Channel/Bray Fault, accommodating some of the major NW–SE dextral strike-slip displacement along that fault (Holder & Leveridge 1986*a*). The relationship of the three Silesian formations has not been closely defined. It is evident

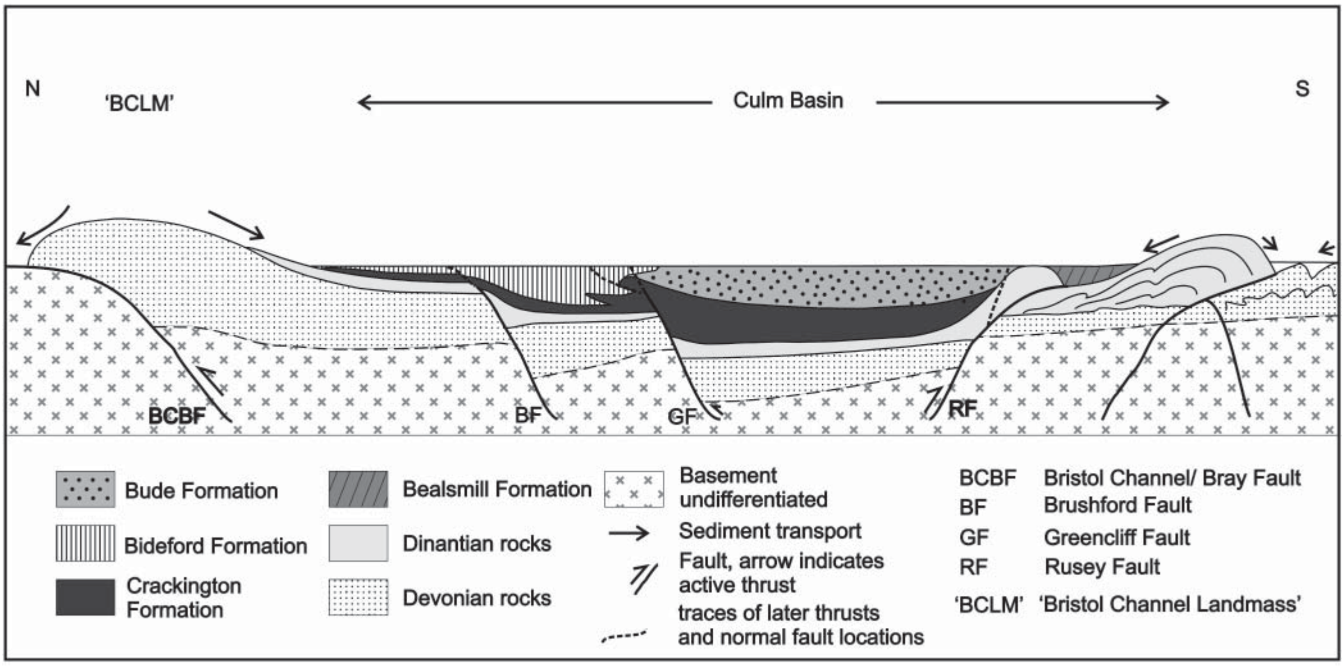


Fig. 10.14. Diagrammatic section of the Culm Basin rift graben during the Early Westphalian, showing the inverted Dinantian marginal sequence to the south, overthrusting on the Bristol Channel/Bray Fault, and progressive inversion on the basin-controlling faults.

that Crackington sedimentation commenced in the inner basin, only overflowing later in the Namurian into the northern sub-basin, where earlier sandstone, as at Venn, may represent a feeder system. The Bideford Formation was largely confined to the restricted northern sub-basin, deltaic sediments accumulating there, whilst Crackington turbidite sedimentation continued in the main basin. The Bude Formation represents the overspilling of the northerly deltaic sediments into the main shallowing central basin. Apart from differing interpretations of depositional setting for the same coastal section, there are a variety of deposits and thus environments recorded across the crop of the Bude Formation. This variation, as the basin became shallow and paralic, would result from such as fluvial input or penecontemporaneous movement on the crossing NW–SE Sticklepath Fault constraining depositional domains.

Major thrust-nappe loading affected only the southern part of the passive margin. Elsewhere D_1 inversion produced significant thrust stacking only in the Looe Basin (Fig. 10.16). Originating as an extensional basin, loading was not a major factor in producing the Culm Basin (see Gayer & Jones 1989), but possibly played a role in maintaining open marine conditions in the South Devon and Tavy basins as they inverted in the late Dinantian. Despite such inversions much of the Culm Basin Silesian sediment has a northerly derivation, thus whilst certainly a foreland basin, because of the province translation, it is not a classical example (Allen *et al.* 1986). Nevertheless, as indicated by Lloyd & Chinnery (2002), basin sedimentation and synsedimentary structures do represent an 'orogenic front propagating into its own foreland'.

Late Variscan deformation and related events in the basins of SW England

Contractional deformation, its continuation. The late Variscan orogenic episode of SW England took place in the Carboniferous when the extensional sub-basins of the Culm Basin began to lock-up during their D_1 inversion. Continuing shortening generated deformation throughout the province. For southern crop Devonian and Dinantian rocks, this was a regional second (D_2) deformation, but, for Silesian sequences, it was the first deformation of consolidated rocks (cf. Lloyd & Chinnery 2002), as it was for the north crop Devonian where deformation phases merged. As a single 'closed' system the province deformed within a short time interval. That episode is constrained by the latest deformed and earliest succeeding undeformed sediments, dating of the metamorphism associated with structure development, and intrusion of the early granite of the province.

The youngest deformed rocks of the Culm Basin are those of the Bude Formation of early Westphalian C age (see above), and the oldest unconformable overlying rocks currently recognized are possibly Late Stephanian (Edwards & Scrivener 1999). The latter is on the basis that a considerable red-bed sequence is mapped beneath a biotite–lamprophyre lava flow dated at 290.8 ± 0.8 Ma (Chesley in Edwards & Scrivener 1999) in the Crediton Trough. The low grade of metamorphism of the deformed Silesian rocks, late diagenetic and low anchizone (Warr *et al.* 1991), has precluded its radiometric dating. A few age-determinations from slates in the north crop Devonian (Dodson & Rex 1971) group around 305 Ma (Warr *et al.* 1991). This compares with dates from a 'Start–Perranporth' zone (315–295 Ma) where secondary transposition fabrics are strongly developed (Holdsworth 1989; Steele 1994). Between the Rusey Fault and Tintagel, some 8 km to the SW, the 'Tintagel High Strain Zone' (Sanderson 1979) has been attributed to tectonic reworking during this deformation by underthrusting of the Culm Basin (Andrews *et al.* 1988; Warr 1989). K–Ar whole-rock dates from the slates of the zone are younger (279–289 Ma: Warr 1991) but the high closing temperatures recorded (Primmer 1985) and extended phase of metamorphism (Warr *et al.* 1991) suggest the overprinting effect of epizonal metamorphism associated with later (Bodmin Moor) granite emplacement and cooling. The oldest granite cupola, Carnmenellis, with an emplacement age of 294 ± 1 Ma (Chesley *et al.* 1993), cuts structures assigned to the regional D_3 deformation (Alexander & Shail 1996). The Bodmin Moor Granite (287 ± 2 Ma: Darbyshire & Shepherd 1985) sourced the extrusive rhyolite lavas of the Rame Peninsula (Leveridge *et al.* 2002) that rest unconformably on deformed Devonian rocks. The deformation thus appears to have been at *c.* 305 ± 5 Ma, the Late Westphalian–Early Stephanian.

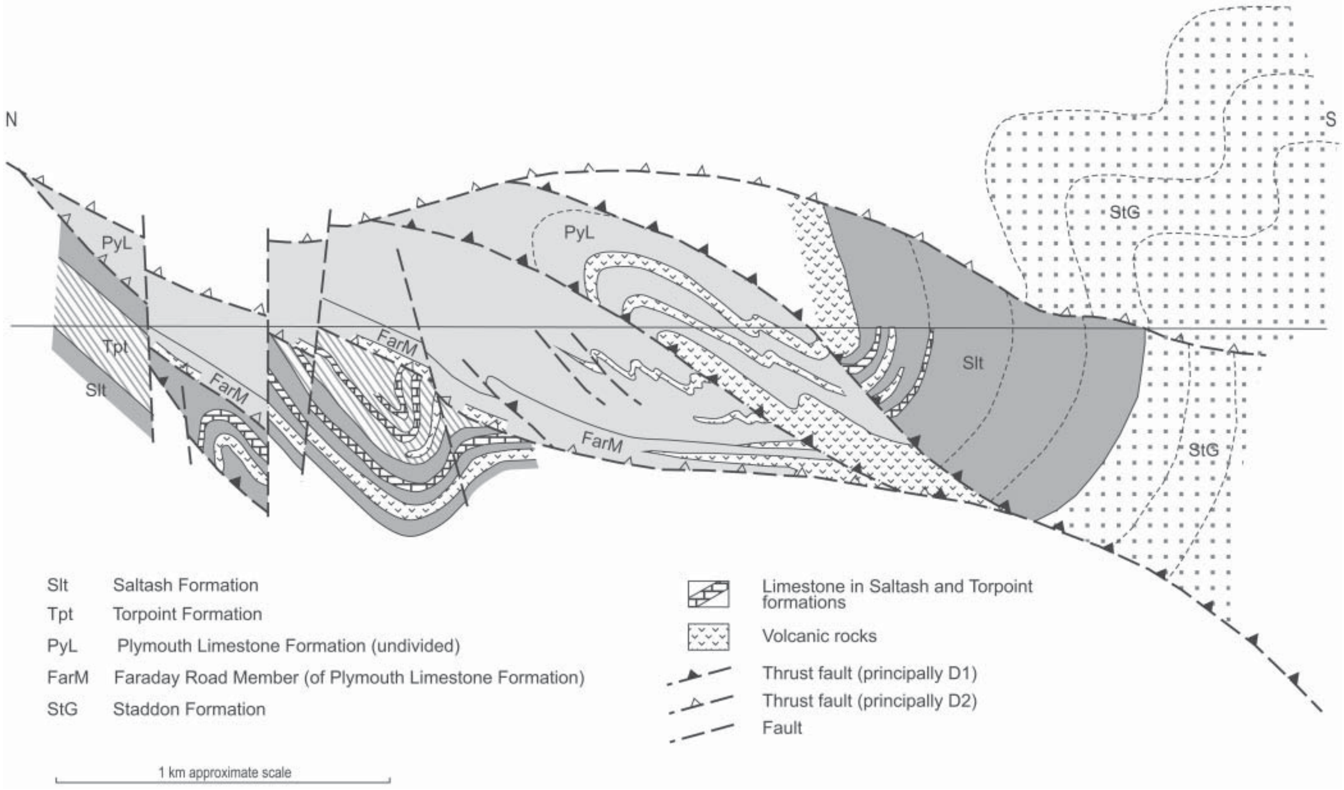


Fig. 10.15. Stuctural section through Plymouth and Plymouth Sound, showing the Looe Basin inversion antiform to the south, and the role of D_1 and D_2 out of sequence thrusting in accommodating the overthrusting of Looe Basin and Plymouth High deposits on to the South Devon Basin sequences to the north. After Leveridge *et al.* (2002).

In the south crop Devonian and Dinantian rocks D_2 structures are not ubiquitous, comprising sporadic zones of mesoscopic folds with associated cleavage, and thrust faults. Minor folds, generally trending between E and NE, verge northwestwards (e.g. Freshney *et al.* 1972; Alexander & Shail 1996). The gently to steeply southerly inclined axial plane cleavage is a crenulation cleavage that generally transposes D_1 fabrics only very locally. However, it can be the predominant fabric more extensively, as on the west Cornwall coast between Perran Sands and Holywell Bay, where an E–W zone some 1.5 km wide has strong transposition and dextral transpressive fabrics (Holdsworth 1989; Steele 1994: BGS unpublished data). Thrust faults are commonly associated with the fold zones, and over much of the subprovince it is the dominant expression of D_2. Thrusts are generally subhorizontal to gently inclined transporting northwards, cutting across early structures (Fig. 10.15), with displacements up to a few kilometres occluding major parts of successions (Leveridge *et al.* 2002, 2003*a*). They are significant structures in the sub-province. Such an out-of-sequence thrust (subsequently rotated to verticality by D_3) juxtaposes the Start Schists with lower grade rocks to the north. On the north Cornwall coast, the Trebetherick Thrust accommodates closer juxtaposition of north- facing (hanging wall) and south-facing (footwall) structures at the 'Padstow Confrontation' (Selwood *et al.* 1998).

D_2 structures are in parts of the province coaxial with D_1 but elsewhere have an anticlockwise divergence by up to 30°. Where there is obliquity with earlier basinal and deformational structures transpressive fabrics are developed (e.g. Andrews 1993; Steele 1994).

The Silesian deposits of the Culm Basin, north of the Rusey Fault, were deformed for the first time during this episode. Although D_2 structures just south of the fault verge northwards (Freshney *et al.* 1972), its northerly dip dictated the southerly vergence of stuctures of its hanging wall along the southern margin of the Silesian fill of the basin during inversion. There is evidence of southerly out-thrusting of Silesian rocks during this process (Warr 1989; Andrews *et al.* 1988) but remnants of the inverted limb of the major inversion antiform are present over an 8 km section north of the Rusey Fault (Freshney *et al.* 1972). South-facing recumbent folding of this zone (Plate 17), becomes generally upright chevron folding in the centre of the basin between Bude and Welcome Mouth, and further north overturning northwards predominates in Silesian (Freshney *et al.* 1979) and North Devon Basin rocks (Edmonds *et al.* 1975). Sanderson (1984) attributed this fan of structural facing to the deformation and closure of a filled Culm Basin graben. A strong spaced cleavage to the south decreases in intensity to the basin centre and then increases northwards (Freshney *et al.* 1979*b*). The slaty cleavage and epizone/high anchizone metamorphism of the pre-Silesian rocks to the north is largely a function of burial (Warr *et al.* 1991) and tectonic loading. This loading, by inverted Silesian rocks, is poorly constrained by lack of modern analysis, but is intimated in the NW area of the Silesian crop where thrust nappes probably represent inversion on earlier basin faults. There the Bideford Formation is bound to the north by a sequence of northerly transporting thrusts (Prentice 1960*a*; Cornford *et al.* 1987), along strike from the Brushford Fault (Fig. 10.13), and to the south by major steep E–W faulting (Burne & Moore 1971), referred to as the Greencliff Fault. Cornford *et al.* (1987) showed, on the basis of vitrinite reflectance, that the Bideford Formation had been buried less deeply than correlative rocks just to the south (4.5/5.5 km rather than 5.7/7.0 km) and proposed juxtaposition by major overthrusting from the south.

At the end of the Westphalian all basins and deposits had been inverted and deformed (Fig. 10.16). It is probable that the inverted Culm Basin formed a substantial edifice, and collision to the south had produced significant thrust nappe loading

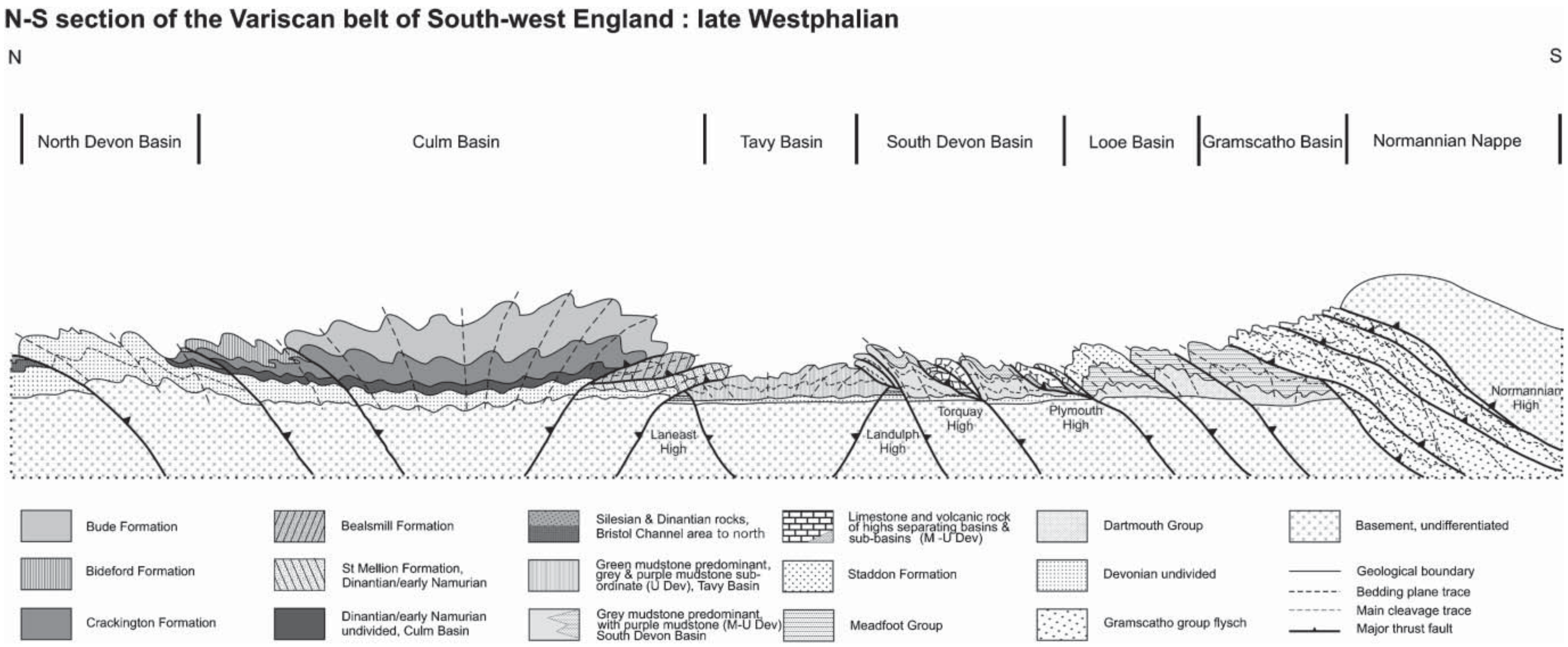


Fig. 10.16. Sketch cross-section of the Variscan belt of SW England, late Westphalian.

of the passive margin, but there does not appear to have been major crustal thickening during the protracted contractional deformation of the province.

Late Variscan orogenic collapse. The final phase of Variscan deformation was active in the brief interval between the inversion of the Silesian Culm basin and the intrusion of the earliest granite cupolas at the time of the Carboniferous–Permian transition. It is the D_3 regional deformation that, in the southern part of the province, represents a reversal of structure vergence (e.g. Alexander & Shail 1996). In south-crop pre-Silesian rocks this deformation is commonly represented by narrow zones of tight–open folds, with associated northerly inclined crenulation cleavage, or minor thrust duplex systems (e.g. Richter 1969; Goode & Leveridge 1991). It is also expressed in rotations of both D_1 and D_2 structures into major southward-verging monoformal/antiformal folds. Attributed to 'backfolding' by Coward & McClay (1983) this is the most notable of these structures with a steeply inclined limb, 1.5 km or more across, extending from north of the Start Peninsula westwards across the peninsula (Leveridge *et al.* 2002) to north of Perranporth (Hobson 1976).

Within the Silesian rocks the fan of chevron folds across the basin is affected by secondary structures. Described by Freshney *et al.* (1979b) as anomalous folds, namely 'box folds, crumples and recumbent folds', those authors ascribed them to either static load under low confining pressure, or for the latter, gravity sliding. Lloyd & Whalley (1986) attributed these structures to the south of Bude to modifications of the earlier chevron folds by a later phase of simple southerly shear. This was designated the D_3 deformation of Silesian rocks by Lloyd & Chinnery (2002), who counted synsedimentary thrusting as D_1 structures. Freshney *et al.* (1979b) noted that these late structures in the northern part of the Silesian crop indicated northerly translation.

Alexander & Shail (1995, 1996) described D_3 in terms of zones of distributed shear, detachments and brittle listric faults, all displaying top-to-the-SE sense of shear in the south of the province. The emplacement of the St Mellion Outlier some 22 km south of the Culm Basin main crop, on flat-lying thrusts (Fig. 10.17), has been attributed to this deformation by Leveridge *et al.* (2002) who invoked post-D_2 gravitational collapse of inverted Culm Basin deposits. This was the early stage of the deformation process that became progressively

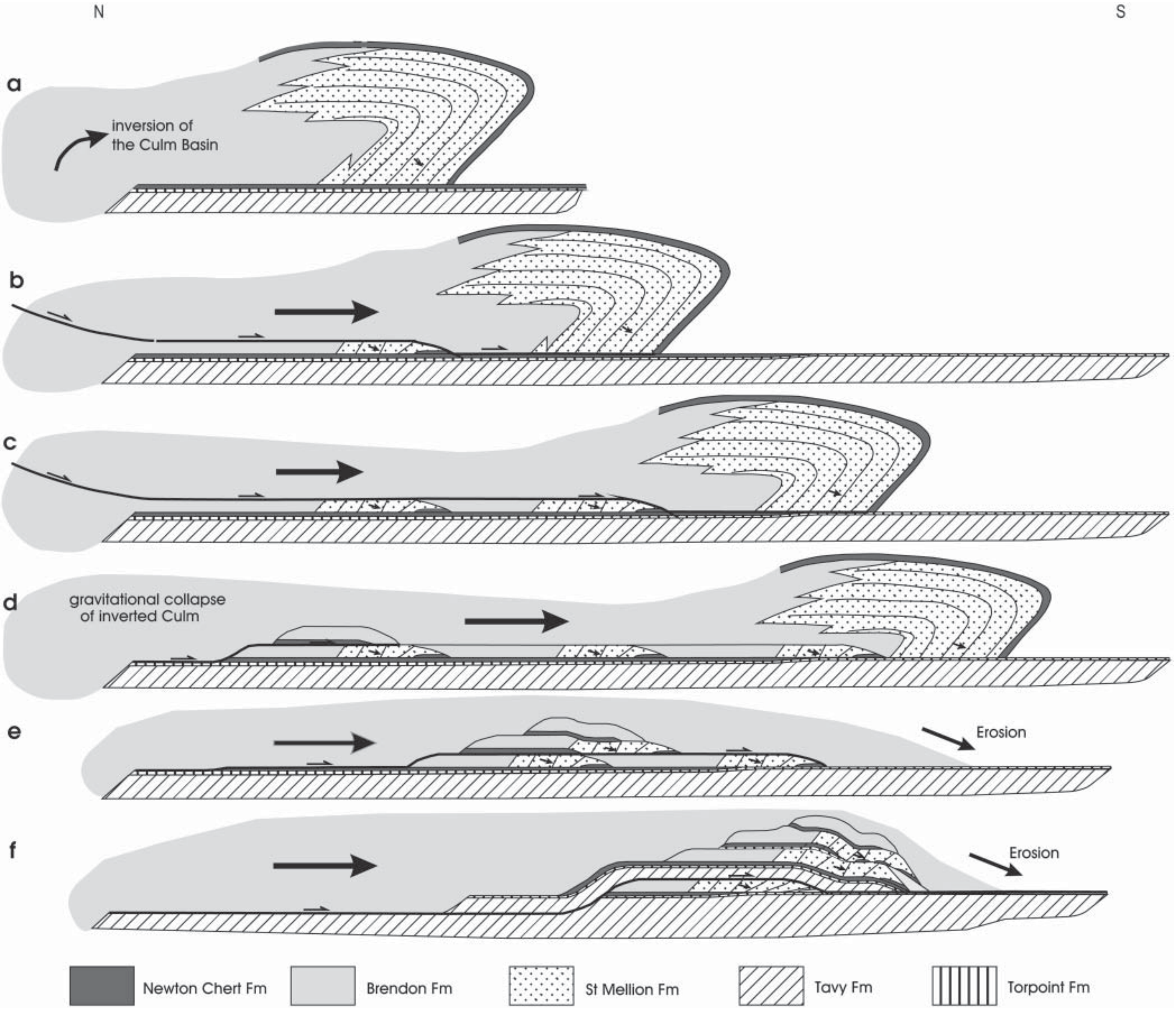


Fig. 10.17. Sequence diagram of the inversion of the southern marginal Culm Basin Dinantian deposits, and subsequent gravitational collapse, which produced the family of thrust slices of inverted rocks of the St Mellion Outlier some 22 km to the south. After Leveridge *et al.* (2002).

more brittle (Alexander & Shail 1996), with extension on earlier thrusts and then development of moderate to steep extensional faults (e.g. Freshney *et al.* 1972; Leveridge *et al.* 2002), some rooting into original basin controlling faults (Fig. 10.14). Continuing into the Permian, with coeval basin formation and granite intrusion at the front of the extending major southerly nappes (Holder & Leveridge 1994; Shail & Wilkinson 1994), it represented orogenic collapse.

The Variscan basin of South Wales

During the Silesian, South Wales comprised one of a series of foreland basins developed in front of the northwardly propagating Variscan orogenic belt. These basins essentially trend E–W and extended from southern England across northern Europe to the Urals. Silesian sediments within South Wales are exposed in a structurally complex WNW–ESE-trending and inwardly plunging synformal structure which extends westwards into SW Dyfed (Fig. 10.18). The structure is asymmetric, the southern limb (the south crop), dips steeply north, and the northern limb (north crop) dips gently south. The succession represents an erosional basin-remnant preserved within the Variscan fold belt. The impact of the Variscan Orogeny on basin development in South Wales is recorded by thickness and lithological changes in the Silesian succession (e.g. Kelling 1988).

Silesian sedimentation in South Wales

The preserved Silesian succession in the South Wales Coalfield and SW Dyfed is up to 3.5 km thick, ranges in age from basal Namurian to early Stephanian (Figs 9.22, 9.26 & 9.27) and was deposited in just over 21 Ma (time scale of Lippolt *et al.* 1984). The stratigraphy and sedimentology of the basin are described in Chapter 9. A brief summary of the sedimentology relevant to foreland basin development is presented here. The succession can be divided into three informal units.

Pendleian–mid-Yeadonian (Cumbriense Marine Band)

Pendleian sediments only conformably overlie Dinantian strata in the basin centre (see Fig. 10.19, 9.22 and Jones 1974 for further details;). Sedimentation was restricted initially to a narrow NE–SW-trending trough located between the Careg Cennan Distrubance (CCD) and the Tawe Disturbance (TD) for the Pendleian and much of the Arnsbergian (Fig. 10.19). The progressive expansion of the basin took place in Chokerian–early Marsdenian times with basinwide deposition by Superbilinguis Marine Band times. By mid-Marsdenian times sedimentation probably extended north of the present day Silesian outcrop (Fig. 10.19).

Pendleian–mid-Marsdenian sediments comprise interbedded quartzitic sandstones, conglomerates and mudstones that record deposition as a series of braid deltas that prograded into a storm-dominated, shallow-marine environment (George 2000).

Lateral facies variations reflect the palaeotopography of the basin. Sandstone deposition occurred intially throughout the narrow NE–SW trending Pendleian trough. However, as the basin continued to subside and/or sea-level rose, deeper water covered the Gower depocentre and shelfal mudstone sedimentation prevailed with sandstone deposition restricted to the basin flanks. Deeper water sedimentation persisted up to the mid-Yeadonian with deposition of inner shelf–offshore transition

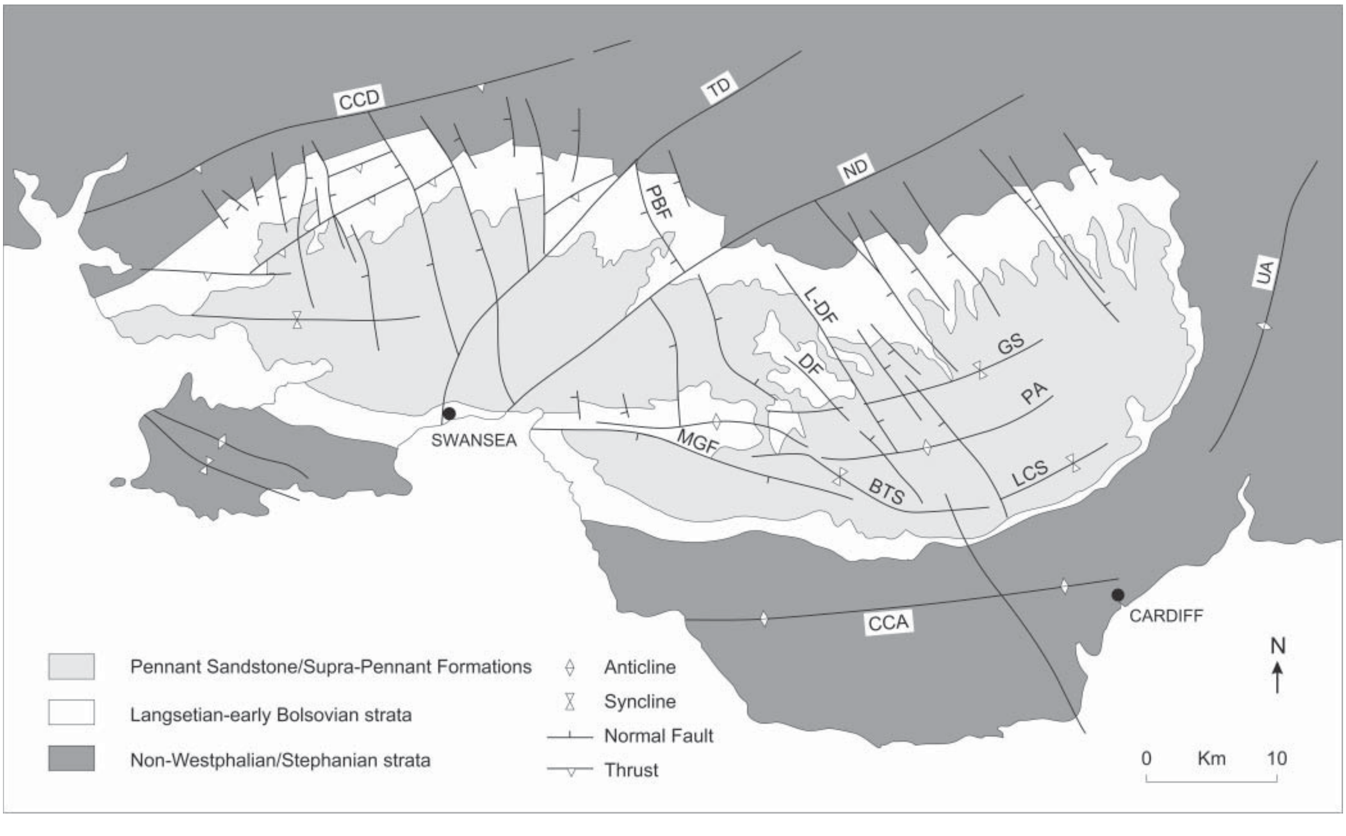


Fig. 10.18. Geological map of the South Wales Coalfield showing the main structures that are thought to have influenced sedimentation during deposition of the Langsettian–early Bolsovian interval. CCD, Carreg Cennan Disturbance; TD, Tawe Disturbance; PBF, Pwllau Bach Fault; L–DF, Llanwonno–Daren-Ddu Fault System; DF, Dinas Fault; MGF, Moel Gilau Fault; PA, Pontypridd Anticline; GS, Gelligaer Syncline; CCA, Cardiff–Cowbridge Anticline; UA, Usk Axis; BTS, Bettws–Tonyrefail Syncline; LCS, Llantwit–Caerphilly Syncline (after Hartley 1993*a*).

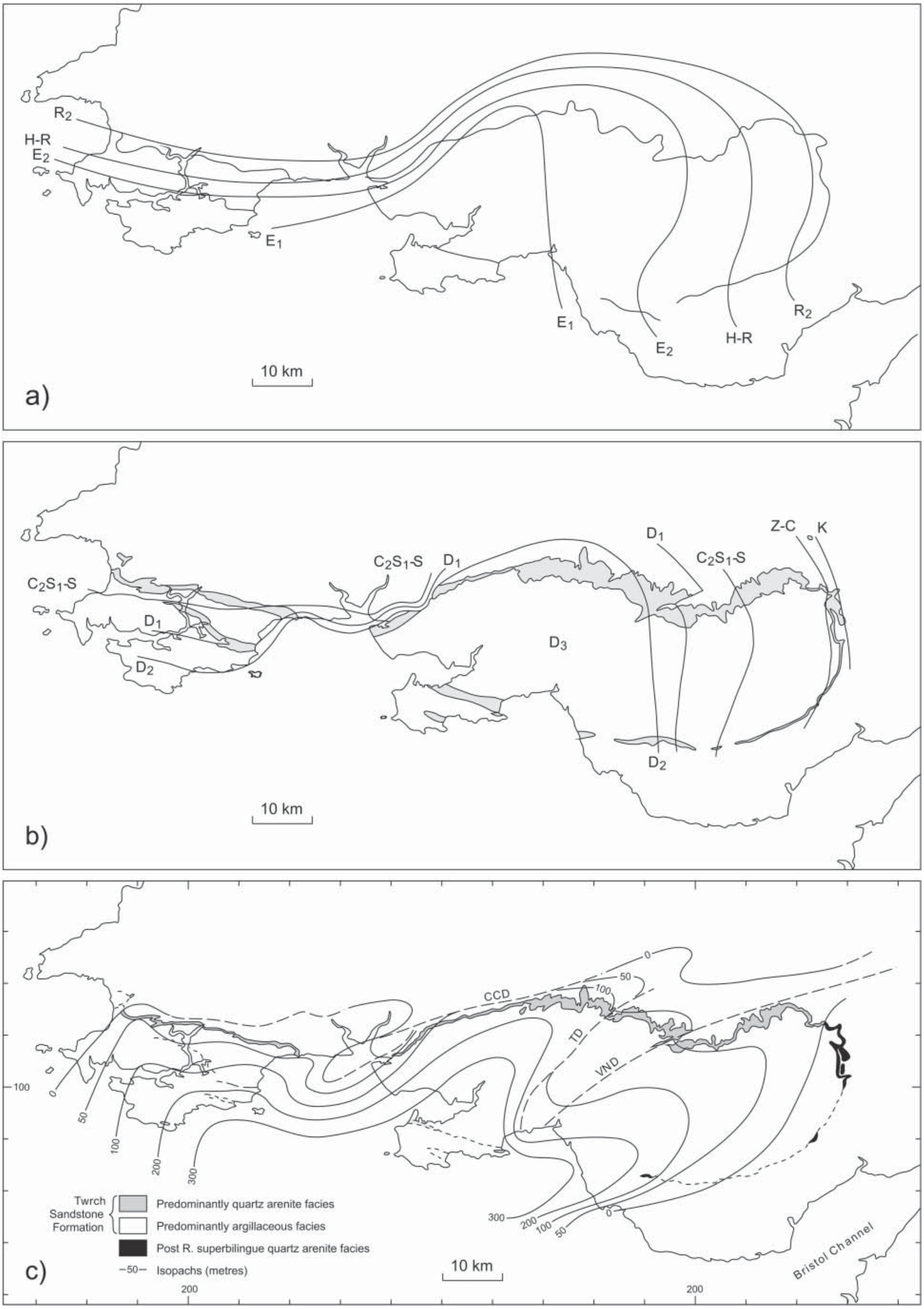


Fig. 10.19. (**a**) Progressive onlap of Namurian strata from the central part of the basin from Jones (1974). (**b**) Subcrop age of the Dinantian strata beneath the Namurian unconformity from George (1970). (**c**) outcrop and isopach map of the Pendleian–mid Marsdenian (*R. superbilinguis*) interval for South Wales, modified from George (2000).

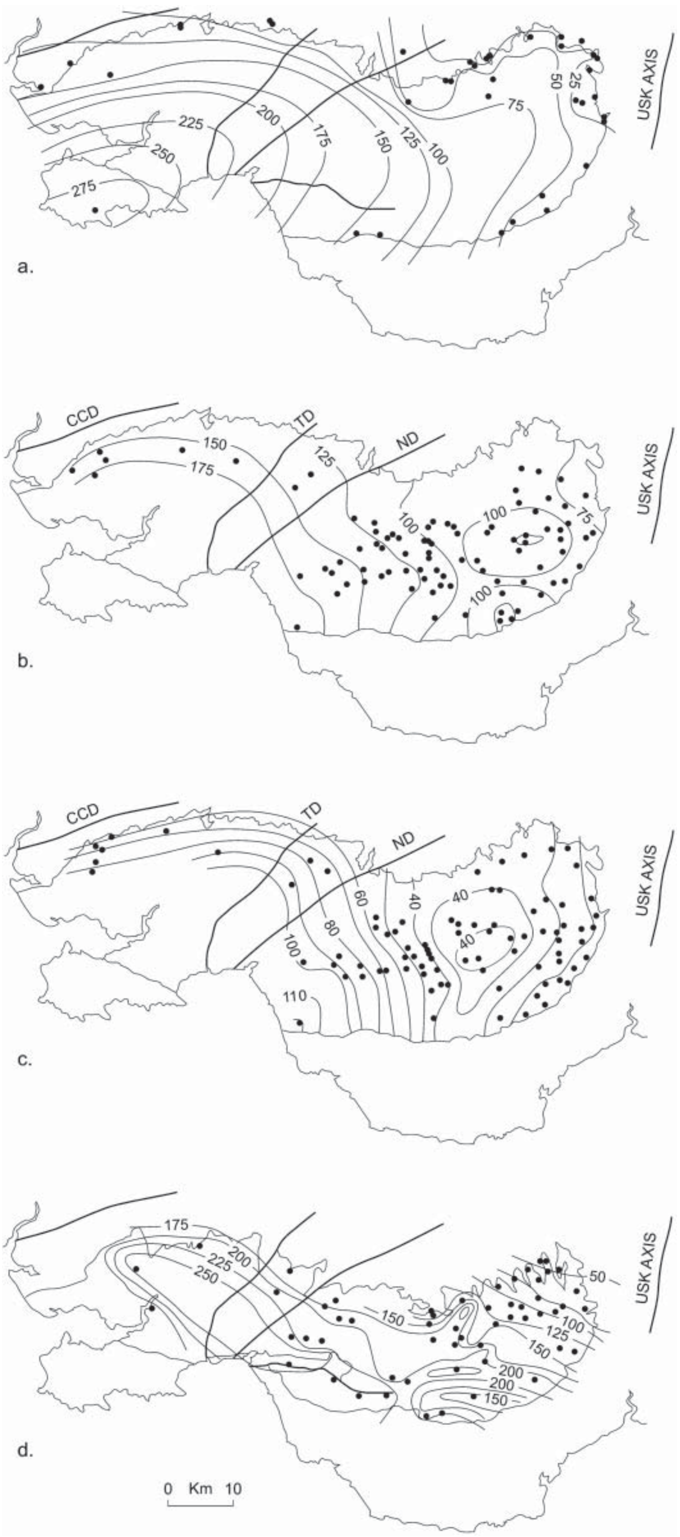


Fig. 10.20. Isopach maps (in metres) for the intervals: (**a**) Mid-Marsdenian–Mid Yeadonian; (**b**) Five Feet/Gellideg to Amman (Vanderbeckei) marine band; (**c**) Amman marine band to Two Feet Nine Coal; (**d**) Brithdir Member of the Pennant Sandstone Formation. Dots refer to control points. Thicknesses corrected for tectonic dip and repetition due to thrust faulting, data from Squirrel & Downing (1964, 1969), Woodland & Evans (1964), Parry (1966), Adams (1967), Archer (1968), Barclay (1989) and unpublished British Coal borehole data.

zone mudstones including marine bands (Fig. 10.20). Occasional fine- to medium-grained sharp-based/incised quartzite bodies are interbedded with the mudstones and represent lowstand shoreline/incised valley-fill deposits (Hampson 1998; George 2000).

Mid-Yeadonian–early Bolsovian

Following deposition of the Cumbriense Marine Band a significant relative sea-level fall resulted in a change from a mainly shelfal environment to a lower coastal plain environment. Coastal plain deposits then dominated the basin-fill succession until the early Bolsovian (Cambriense or Upper Cwmgors Marine Band; Figs 9.22 & 9.27). Both upper and lower coastal plain deposits can be recognized. Lower coastal plain deposits occur from mid-Yeadonian to mid-Langsettian and late/Duckmantian–early Bolsovian times. They are characterized by horizontally laminated mudstones with marine bands, erosive-based, medium-coarse grained, cross-stratified quartzitic sandstone bodies, rippled siltstones, fine-grained sheet sandstone beds and thin coal seams (Jones 1989*a*, *b*; Hartley 1993*a*, *b*). Quartzite-bodies are interpreted as low-sinuosity fluvial deposits some of which form incised valley fills (Hampson 1998; George 2001). Upper coastal plain deposition prevailed in mid-Langsettian–late Duckmantian times and is characterized by well developed 5–30 m thick, mudstone-dominated coarsening upwards cycles containing only brackish water fauna. Mudstones grade through rippled siltstones to medium-grained sheet sandstones. Sandstones have rootleted tops and are overlain by thick coal seams and mudstones with a brackish fauna. Occasional siltstone or coarse-grained sandstone channel-fill units are present. Basin-wide coal seams overlain by brackish water fauna indicate the termination of peat accumulation by basin-wide flooding that resulted in lake development. Re-establishment of the clastic supply infilled lakes through overbank flood events (sheet sandstones) supplied by low-sinuosity fluvial systems (Hartley 1993*a*).

Early Bolsovian–early Stephanian

Following deposition of the Upper Cwmgors/Cambriense Marine Band (Fig. 9.26) a marked change in sedimentation took place. The preceding lower coastal plain succession was replaced by lithic sandstones of the Pennant and Supra-Pennant Sandstone formations. The Pennant Sandstone Formation is dominated by thick, medium- to coarse-grained, locally pebbly, channelized sandstone bodies interbedded with occasional horizontally laminated mudstones, cross-laminated siltstones and coals. They are interpreted to represent large-scale, bedload-dominated, braided fluvial systems deposited on an alluvial braidplain (for further details see Kelling 1968, 1974; Jones 1989*a*, *b*; Jones & Hartley 1993; Hartley 1993*b*). The mudstone-dominated Supra-Pennant Formation indicates a change to a more floodplain-dominated high-sinuosity fluvial succession. Sedimentation continued into the Stephanian although the top of the basin-fill sequence is truncated.

Subsidence patterns and rates

Prior to the onset of Silesian sedimentation in South Wales, regional deformation resulted in uplift and exposure of the underlying Dinantian carbonate succession. This is revealed by the erosion and karstification of the Dinanatian carbonates and the subcrop map of Dinantian strata beneath the Namurian (Fig. 10.19). This map shows the progressive onlap of Namurian strata northwards, eastwards and westwards from a basin depocentre located close to the Gower (Ware 1939; George 1956). Isopach maps for the Dinantian succession show regional thickening to the south of Pembrokeshire into the present day Bristol Channel area (George 1974), such that the basin depocentre narrowed and shifted to the NE from late Dinantian to early Namurian times. The subcrop map also illustrates the extent and location of major uplift axes which coincide with major lineaments notably the Carreg Cennan Disturbance, the Tawe Disturbance and Usk Axis, suggesting that movement on these structures was responsible for uplift and basin reorganization during late Dinantian–early Namurian times.

Further evidence for continued tectonic activity during sedimentation is provided by isopach and subcrop maps for selected stratigraphic intervals (Figs 10.19 & 10.20). These show a number of common characteristics, including:

(1) The NE–SW oriented trough between the Swansea/Gower area and Ammanford established in the early Namurian formed the basin centre throughout the Silesian.
(2) Thicknesses decreased markedly away from the depocentre particularly to the east and more gradually to the north and west.
(3) Isopach maps for all time intervals up to the Brithdir Member can be seen to parallel the Usk Axis to the east and the Carreg Cennan Disturbance to the NW of the Coalfield. It is likely therefore that these features were active during sedimentation particularly as rapid thickness changes occur adjacent to these structures. The Neath Disturbance may have also acted as a positive area as Westphalian sediments thin across it.
(4) The isopach maps suggest that there is a northwards shift of the basin depocentre (to just north of Swansea) as indicated by the closure of isopachs around the southern basin margin by the end of deposition of the Brithdir Member of the Pennant Sandstone Formation. This supports the observations of Kelling (1988).

Localized variations in thickness are common and are considered to be related to synsedimentary tectonism (Jones 1989*a*; Hartley & Gillespie 1990; Hartley 1993*a*). In particular, patterns of seam-splits, thickness variations and channel orientation in Langsettian and Duckmantian upper coastal plain deposits indicate tectonic control on sedimentation by a number of active structures. Thinning occurs over E–W-trending structures such as the Pontypridd Anticline with thickening into synclines such as the Bettws-Tonyrefail, Gelligaer and Caerphilly synclines; these growth folds have been ascribed to the reactivation of basement structures during early Variscan compression (Jones 1989*a*; Hartley 1993*a*). A number of NW–SE-trending faults are thought to have influenced sedimentation by controlling the position of seam-splits (Hartley 1993*a*). These faults show normal displacement at the present day, but are considered to represent originally reverse-fault bounded blocks that have been subsequently reactivated as extensional faults (Jones 1991; Hartley 1993*a*). The presence of soft-sediment deformation and gravity slides with channel-fill deposits preferentially developed on footwall blocks also indirectly suggests that tectonic activity was active during sedimentation (Hartley & Gillespie 1990).

Subsidence curves for the Silesian section in South Wales have been presented by Kelling (1988) and Burgess & Gayer (2000). They indicate relatively slow rates of subsidence through the Namurian, with a rapid increase at or shortly after the Namurian–Langsettian boundary (Fig. 10.21). The relatively rapid subsidence rates persisted throughout the Silesian, with no significant change associated with onset of Pennant Sandstone Formation deposition. Both sets of authors also indicate a northward migration of a depocentre through the Silesian.

Drainage systems and provenance

Palaeocurrent data for the Silesian succession has been published by a number of authors (Bluck 1961; Bluck & Kelling 1963; Thomas 1967; Kelling 1968, 1974; Williams 1968; George 1970, 1982, 2000, 2001; George & Kelling 1982; Jones 1989*a*, *b*; Hartley 1993*b*; Hampson 1998). Namurian sandstones were derived from the north, east, west and SE basin margins with an open marine connection south of the Gower area. A similar scenario prevailed throughout the Langsettian–late Duckmantian, with coarse-grained, lithic sandstones derived from the south whilst finer-grained channel-fill units were sourced from the east, north and west. Studies of late Duckmantian–early Bolsovian fluvial systems indicate that coarse-grained lithic sandstones were largely derived from the south and to a lesser extent from the east.

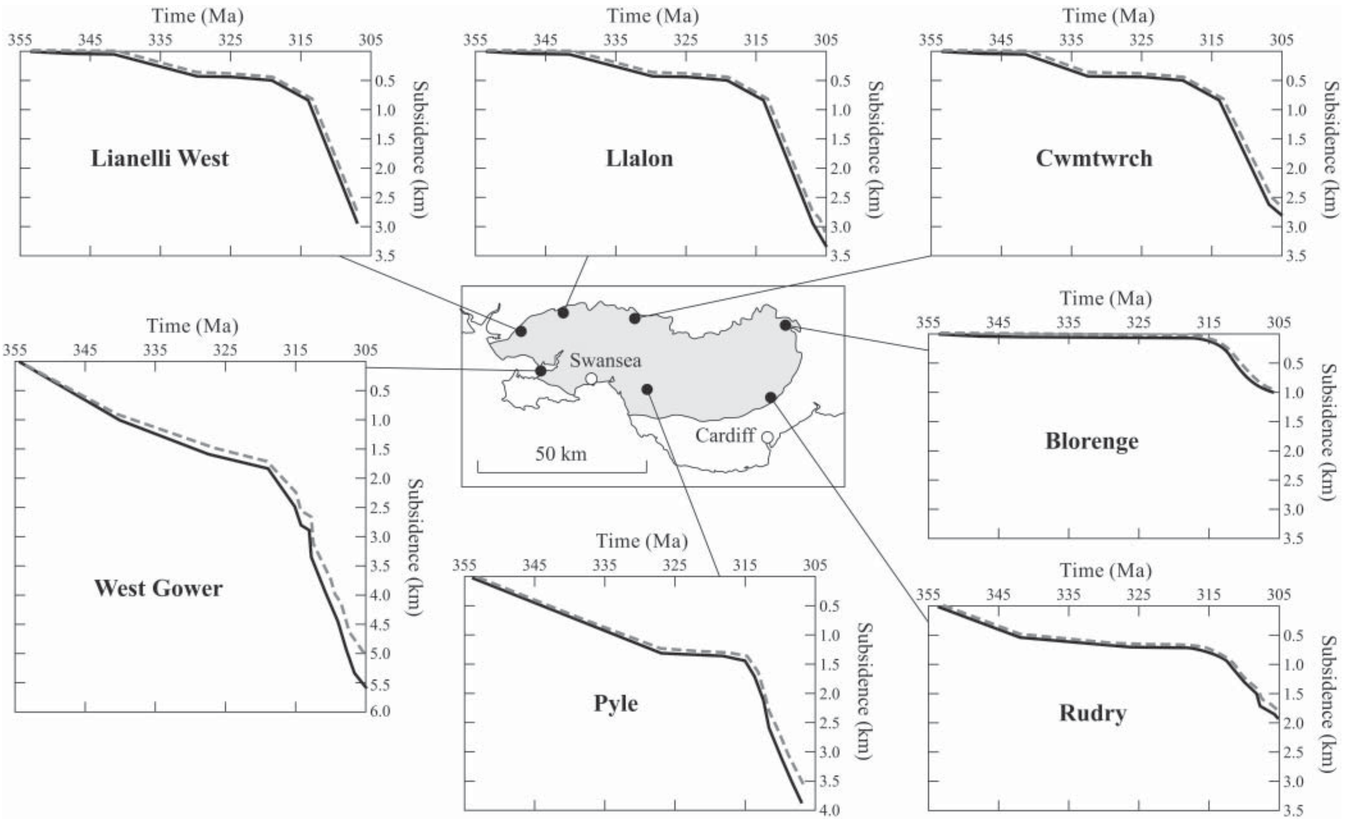


Fig. 10.21. Backstripped subsidence curves, solid line is total subsidence and dashed line is tectonic subsidence, from Burgess & Gayer (2000).

These palaeocurrent data derived from direct outcrop measurements suggest a predominantly centripetal drainage system throughout the Namurian, Langsettian, Duckmantian and into the early Bolsovian. Sediment derived from the basin flanks was transported towards the south/SW towards the marine connection and basin depocentre. This interpretation is in direct contrast to the NW–SE drainage direction proposed by Rippon (1996) from coal seam washout orientations in late Langsettian and Duckmantian strata in the east of the South Wales Coalfield. Rippon based his interpretation on washout trends and presented no palaeocurrent data. It seems unlikely that fluvial channels would effectively drain up-dip, away from the basin depocentre towards the basin margin. Unfortunately, the drainage pattern proposed by Rippon has been adopted by other authors (e.g. Guion *et al.* 2000).

Palaeocurrent and provenance studies of the Pennant Measures indicate derivation from a predominantly southerly source area composed of immature lithic detritus of possible Devonian–Upper Carboniferous age (Heard 1922). Limited quartzitic detritus was supplied to the basin from the east during deposition of the Llynfi and Rhondda Members (Kelling 1974). Abundant coal clasts in channel lag deposits of the Rhondda Member of the Pennant Sandstone Formation were derived from Langsettian coal seams located south of the present-day coalfield where they may have been uplifted by thrusting (Gayer & Pesek 1992). The influx of coarse-grained detritus from the south indicates rapid uplift of a southerly source area.

Synthesis of the South Wales Silesian basin-fill

The Silesian succession in South Wales was deposited during the northward propagation of the Variscan orogenic belt. The foreland basin interpretation is supported by:

(1) The progressive northward migration of the basin depocentre (e.g. Kelling 1988).
(2) A progressive increase in subsidence rates (particularly from the late Namurian early Westphalian onwards), suggestive of increased lithospheric flexure (e.g. Kelling 1988; Burgess & Gayer 2000).
(3) The general east–west orientation of the basin parallel to the inferred Variscan thrust front.
(4) The stratigraphic evolution of the basin-fill from Dinantian carbonates, Namurian shallow-marine and muddy shelfal deposits, through to coal-bearing coastal plain deposits and coarse-grained immature, high-energy braided fluvial systems is typical of the evolution of a foreland basins in general (e.g. the Alps and Pyrenees) and particularly for Variscan foreland basins (Hartley & Otava 2001).
(5) Coal clasts in channel lag deposits within the Rhondda Member of the Pennant Sandstone Formation were derived from erosion of Langsettian coal seams to the south of the present day coalfield indicating rapid burial and uplift (within 5 Ma) probably by thrusting (Gayer & Pesek 1992).
(6) Evidence from the inferred orogenic load in SW England for mid-Namurian age thrusting (Dodson & Rex 1971; Burgess & Gayer 2000).

The influence of the orogenic load on foreland basin subsidence and sedimentological development is interesting. Following uplift and reorganization of the basin in the late Dinantian, relatively low subsidence rates persisted throughout much of the Namurian and coincided with a general deepening of the basin suggesting that subsidence outpaced sediment supply. The main increase in subsidence rate in the basin occurred in late Namurian–early Westphlian times and coincided with a relative shallowing of the basin-fill succession suggesting a possible relationship between enhanced subsidence and increased sediment supply. Rapid subsidence during the remainder of the

Westphalian coincided initially with low-lying coastal plain sedimentation, suggesting that sediment supply kept pace with the increased rate of subsidence at least until the early Bolsovian. The influx of lithic sandstones from a southerly source in the early Bolsovian (Pennant Sandstone Formation), does not coincide with an increase in subsidence, however, the basin-fill changes rapidly from coastal to alluvial plain deposition suggesting that sediment supply outpaced subsidence from Bolsovian to Stephanian times.

Subsidence patterns were also strongly influenced by a number of major long-lived lineaments the most important of which appear to have been the Carreg Cennan Disturbance and the Usk Axis, which controlled the location of the NW and eastern basin margins, respectively. The Tawe and Neath Disturbances appear to have had some influence on sedimentation although not as marked as the other major structures. On a smaller-scale other structures (e.g. E–W-trending anticlines and synclines, NW–SE-trending faults, e.g. Jones 1989*a*, *b*; Hartley 1993*a*) were also active during sedimentation locally controlling channel courses, seam splits and thicknesses.

The marked cyclicity present within the Pendleian–early Bolsovian part of the basin-fill has been attributed to sea-level change driven by glacio-eustatic fluctuations (e.g. Ramsbottom 1978; Hartley 1993*a*; Hampson 1998; George 2000, 2001). This is supported by the basin-wide nature of many of the cycles, although, cycle development was locally affected by synsedimentary faulting and folding in places.

Post-Depositonal Thermal and Structural Development

Crustal thickening across SW Britain led to the progressive lowering of geothermal gradients. Estimates of 40–50 °C km^{-1} have been suggested for the South Wales Coalfield, which may reflect the high degree of crustal thinning that occurred prior to collision (Warr 2000). Although the degree of coalification generally increases with burial depth the NW part of the basin does show an anomalous increase in temperature as indicated by the presence of anthracite grade coal. It is possible that this thermal anomaly reflects a focal point for the upward migration of hotter fluids, driven in fort of the advancing thrust wedge. These fluids infiltrated the coal seams along brittle fractures, causing hydrothermal mineralization and the precipitation of gold before the onset of seam-parallel thrusting (Gayer *et al.* 1998).

The South Wales area forms part of the foreland to the main Variscan deformation belt (represented by the Rhenohercynian zone in SW England). The Variscan deformation front is considered to lie just south of the South Wales Basin (Fig. 9.4). The main phase of deformation within South Wales took place following deposition of the Supra Pennant Formation in the late Stephanian–early Permian. Evidence suggests that deformation of the Silesian succession varied both stratigraphically and geographically (Jones 1989*a*). The amount of deformation and coal rank increase to the west. The relatively incompetent Langsettian and Duckmantian (Productive Measures) strata are the most intensely deformed. The more competent Pennant Sandstone Formation is much less deformed (approximately 6% shortening, with just broad, gentle folds), and acted as a passive roof duplex to an underthrust wedge. Deformation in the coal-bearing succession is dominated by WNW–ESE striking thrusts and folds which produced approximately 30% shortening in the east of the coalfield (Jones 1991) locally rising to 50–67% in the west (Frodsham & Gayer 1997). Structures verge to the north in the north crop with an important south-verging back-thrust zone forming the south crop. Regionally the thrusts propagate to the north in a piggy-back sequence but locally a variety of complex structures evolved, controlled by the position and geometry of coal seams. Thrusts initially moved along the floors of the coal horizons until they locked up and produced hangingwall folds. Further displacements eventually caused the faults to break through the fold hinges, and to initiate new break-back thrusts in either the hangingwall or footwall segments leading to numerous repetitions of coal seams (Frodsham *et al.* 1993; Frodsham & Gayer 1997). The structures are compartmentalized by NW–SE cross-faults.