Tectonics and sedimentation in the Lower Palaeozoic back-arc basin of S. Wales, U.K.: some quantitative aspects of basin development

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Basinal subsidence history, constructed after corrections of sediment thickness for tectonic strain, suggests Arenig stretching by a lithospheric dyking mechanism followed by thermal subsidence from Llandeilo to Pridoli and load-induced flexural subsidence in the Lower Devonian. Sedimentary studies across the basin margin faults augmented by palaeo-waterdepth estimates from volcanic evidence allow quantitative refinement of this conclusion. Consideration of likely crustal and lithospheric thicknesses, constrained by analysis of Middle Devonian basin inversion to a present-day 35 km crust, suggests an initial stretching factor of about 2 with a normal crustal thickness and a somewhat attenuated lithospheric thickness adjacent to the volcanic arc to the NW.

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The Welsh Caldeonides comprise a very thick sequence of late Pre-Cambrian to early Devonian strata orogenically deformed in the middle Devonian. Over the past few years a back-arc basin model for the rapidly accumulated Ordovician–Silurian deposits of the Welsh Caledonides has become increasingly accepted (e.g. Mitchell 1984 and references therein). At the same time increasing emphasis has been put on quantitative evaluations of basin development in a plate-tectonics context (e.g. Dewey 1982). Where a basin fill is only mildly deformed and is geometrically closely defined by seismic mapping, as in the late Mesozoic/Tertiary basins underlying much of the NW European continental shelf, there is reasonable confidence in the results of such analyses. Where the basin fill exhibits later penetrative deformation in a transpressive regime, as do the Welsh Caledonides, the value of such analysis is less obvious: the cumulative errors induced by incomplete three-dimensional control, variable bulk strain effects and ambiguity in radiometric ages possibly precluding the elimination of any alternative models for basin information. This paper argues that a cautious quantitative treatment of the Welsh Caledonides in terms of stretching models is justified and that useful insights can be gained in attempting this.

Basin geometry

The Welsh back-arc basin (Fig. 1) may be thought of as bounded by positive areas in Anglesey and the Irish Sea to the NW and in the Welsh Borderlands and Brecon Beacons to the E and SE (Phillips et al. 1976; Woodcock 1984b). The marginal positive areas are separated from the basin by major fault zones along the Menai Straits to the NW and by the Church Stretton fault to the SE. The back-arc basin is internally divided by the line of the Bala fault into two mega-blocks which display considerable differences in their evolution and which, at least until the Llandovery, cannot be assumed to have occupied their current positions relative to each other. The southern of these mega-blocks is termed the S Wales basin in this paper.

The N Wales mega-block shows geochemical evidence of transition from a volcanic arc to the NW in Anglesey to an extensional basin in the SE during the Caradoc (Leat et al. 1986). It appears to have suffered a much more heterogeneous penetrative deformation than the basin in S Wales (Coward & Siddans 1979). Between the Caradoc and late Llandovery it was a relatively positive area with respect to the basin S of the Bala fault, probably by virtue of being
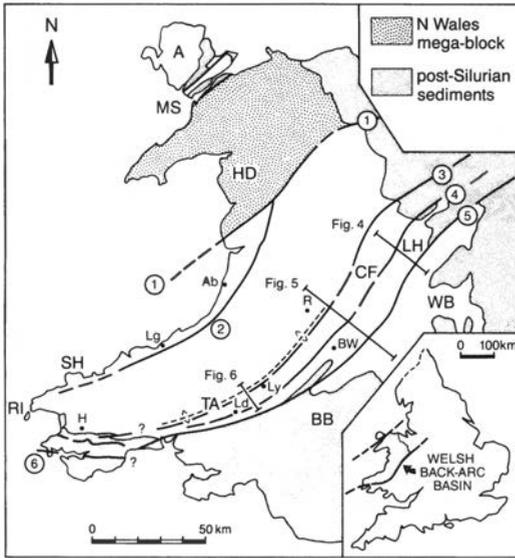
Fig. 1. Location map for the S Wales Lower Palaeozoic basin. Key to tectonic *lineaments*: 1, Bala Fault; 2, Corris–Llangranog lineament; 3, Severn–Tywi lineament; 4, Pontesford lineament; 5, Church Stretton fault; 6, Ritec fault. Key to *localities/towns*: A, Anglesey; Ab, Aberystwyth; BB, Brecon Beacons; BW, Builth Wells; CF, Clun Forest; H, Haverfordwest; HD, Harlech Dome; Ld, Llandeilo; Lg, Llangranog; LH, Longmynd Horst; Ly, Llandovery; MS, Menai Straits; R, Rhayader; RI, Ramsey Island; SH, Strumble Head; TA, Tywi Axis; WB, Welsh Borderland.

sey Island and Aberystwyth display a similarly continuous record from the Arenig to the Llandovery and probably lay close to the bathymetric axis of the basin for much of that time. The southwestward deepening of Ashgill–Caradoc depositional environments towards this basin axis south of the Bala fault together with palaeo-waterdepth estimates have been documented by James (1985). Only in S Wales is there continuity of stratigraphic record into the Devonian. In view of the above facts the following quantitative analysis of the development of the back-arc basin is based on data from the S Wales mega-block. A synopsis of the basinal infill therein is given in Fig. 2.

Data sources

Stratigraphic data for this analysis are drawn from Williams et al. (1972) for the Ordovician, Cocks et al. (1971) for the Silurian and Turnbridge (1986) for the Devonian. Important recent review papers giving extensive introduction to the literature are those of Woodcock (1984b) dealing with sedimentation and tectonics are Bassett (1984) which in multidisciplinary with a strong stratigraphic content. Recent work on volcanism is summarized by Stillman & Francis (1979).
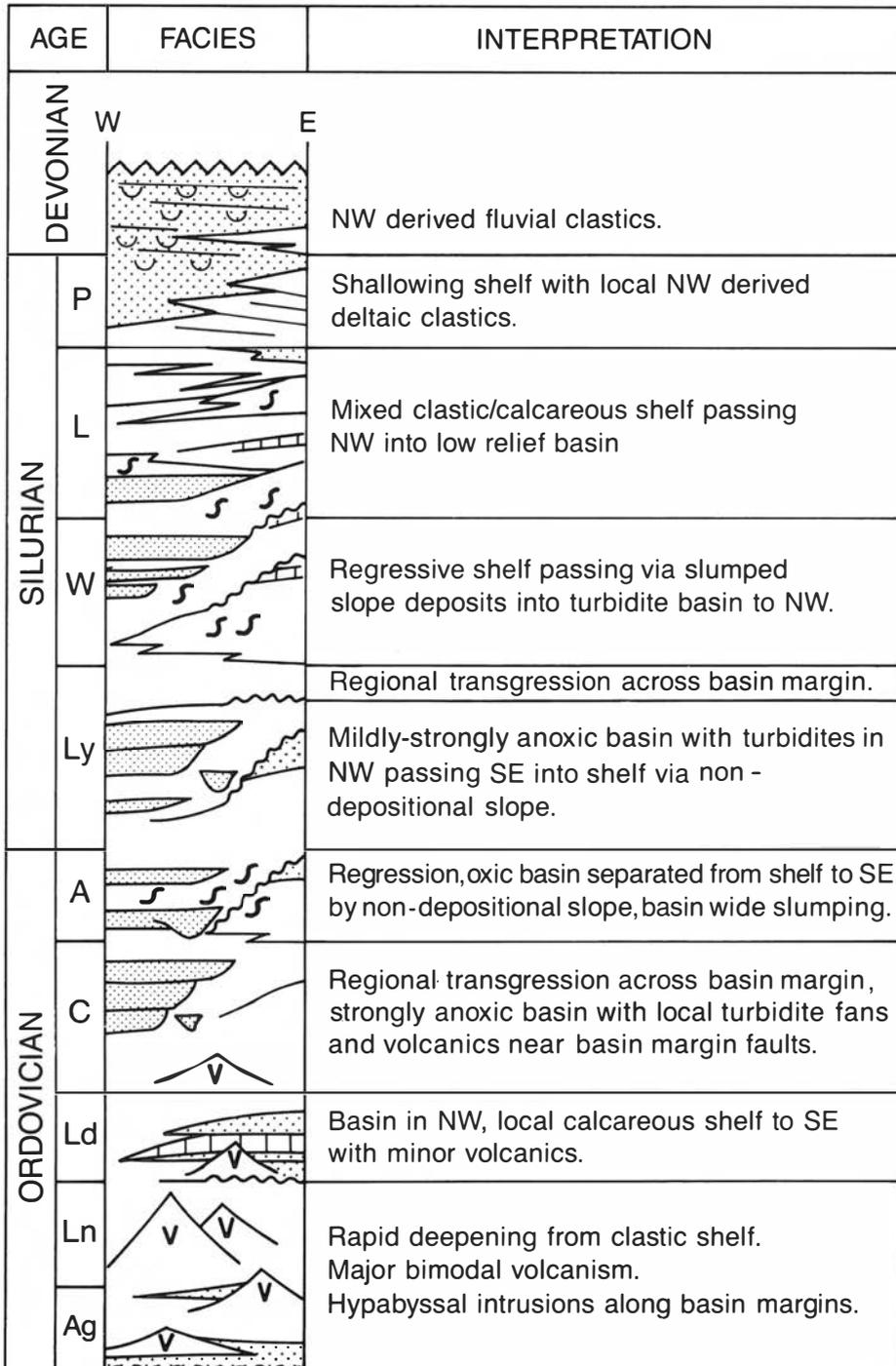
warmer due to strong Arenig–Caradoc volcanic activity although comparable thickness of Caradoc volcanics and sediments combined are preserved either side of the Bala fault in the NE of the basin (Rast 1969:329). In the Arenig–Llandeilo the mega-block was probably a negative area with respect to the S Wales basin due to rapid loading by thick volcanic piles near the volcanic arc to the NW. The above analysis of mega-block relationships assumes limited (?less than 10 km) pre-Llandovery strike-slip movement on the Bala fault.

The major Cambrian developments of the Harlech Dome in comparison with much thinner time-equivalent sequences in SW Wales hint at considerable differences between the mega-blocks in the nature of their pre-Arenig foundations.

The S Wales mega-block, in contrast to the N Wales mega-block, shows a rather continuous record of basin-margin evolution between the Tywi axis and the Church Stretton fault (Fig. 1). Moreover, the coastal exposures between Ram-

Quantitative analysis

Favourable aspects

The stratigraphy, facies and thickness of the sediments and volcanic rocks comprising the basin fill have been rather firmly established over many years of study and estimates of palaeo-waterdepth can be made for some intervals (see below). The inception of the basin is accurately known from volcanic evidence and control on origin is aided by geochemical evidence.

The basin fill is deformed by folding and development of cleavage, but the latter is not intense and the higher stratigraphic levels near the basin margin are locally non-cleaved. These sequences are in structural continuity with stratigraphically deeper cleaved sequences and this suggests a

Fig. 2. Simplified stratigraphic overview of the Ordovician–L. Devonian basin fill.

AGE		FACIES	INTERPRETATION
DEVONIAN		W E	NW derived fluvial clastics.
SILURIAN	P		Shallowing shelf with local NW derived deltaic clastics.
	L		Mixed clastic/calcareous shelf passing NW into low relief basin
	W		Regressive shelf passing via slumped slope deposits into turbidite basin to NW.
			Regional transgression across basin margin.
	Ly		Mildly-strongly anoxic basin with turbidites in NW passing SE into shelf via non-depositional slope.
ORDOVICIAN	A		Regression, oxic basin separated from shelf to SE by non-depositional slope, basin wide slumping.
	C		Regional transgression across basin margin, strongly anoxic basin with local turbidite fans and volcanics near basin margin faults.
	Ld		Basin in NW, local calcareous shelf to SE with minor volcanics.
	Ln		Rapid deepening from clastic shelf. Major bimodal volcanism. Hypabyssal intrusions along basin margins.
	Ag		

5 km

Legend:

- Shallow water ⎱ Arenites
- Deep water ⎰
- 𝒮 𝒮 Slumps
- ⊞ Limestones
- V V Volcanics
- 〰 Unconformity

mean overall shortening of about 30%, close to the shortening at which slatey cleavage develops. Moreover, no major strain discontinuities are known. Any reasonable estimate of shortening across the whole basin implies that its original width was probably greater than that of the thickness of underlying lithosphere. The magnitude of cooling effects attributable to finite stretching time and lateral heat flow (Cochran 1983) is thus not likely to be important and analysis in terms of McKenzie's (1978) instantaneous stretching model is thus justified (Jarvis 1984).

Unfavourable aspects

Owing to incomplete exposure, it is obviously impossible to make a cumulative thickness profile for the total basin fill at any one locality or indeed to identify the former positions of the basin axis with confidence.

The pre-stretching thicknesses of crust and lithosphere are unknown and an assumption that both are equally thinned during extension may not be justified (Hellinger & Sclater 1983).

Removal of bulk strain is subject to several uncertainties and the basic data required are very sparse in S Wales. The assumption of plane strain is probably not a serious mistake and the effects of rotational shear strains which result in over-estimation of true strain (Dunnet & Siddans 1971) are probably localized and unlikely to distort regional thickness estimates. The relative roles and timing of buckling and flattening are not known and it may not be safe to assume that tectonic thickening is everywhere maximal or even that thinning may not locally have occurred (Rast 1969:321–323).

Discussion, construction of subsidence history

Thickness variability has been checked as far as possible from the literature to ensure that laterally shifting depocentres do not give a false picture of accumulation rates. The potential problem is not thought to be serious on the scale of, for example, the entire Llandovery, where a well-defined trough, largely axially filled, was in existence (Fig. 5). Migration of the trough axis with time (Woodcock 1984b:331) is very difficult to prove, although eastward migration of the margin of

deep-water sands in the Silurian is well documented. If present, such migration will reduce the potential underestimate of late Silurian thicknesses that has been made. Cumulative present-day thicknesses are plotted as curve *a* in Fig. 3 using a time-scale based on Spjeldnaes (1978) and Bassett (1984) with minor emendation.

Strain data (Coward & Siddans 1979; Rast 1969) are sparse and strain intensity probably increases from the marginal to the axial portions of the basin irrespective of stratigraphic level. Since the thicknesses of the upper part of the preserved succession are from marginal areas and may be somewhat low with respect to those in the axis, the lower strain there in fact makes for greater consistency in the estimation of original thicknesses. Based on a broad-brush estimate of the proportion of competent lithologies (volcanics, thick greywackes) and cleavage intensity, the Arenig–Pridoli portions of the basin fill are assumed to have suffered average layer-parallel horizontal shortening of between 31%
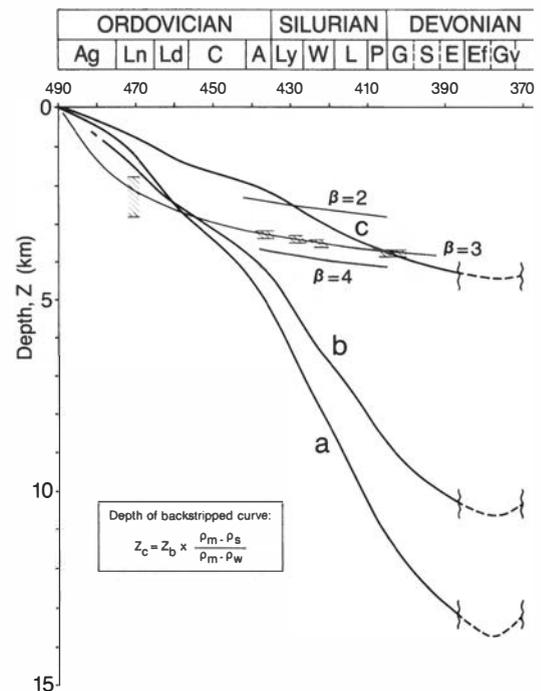

Fig. 3. Subsidence history for the S Wales Lower Palaeozoic basin. (a) Raw data, (b) corrected for tectonic extension, (c) back-stripped. Also shown are palaeo-waterdepth estimates (ranges, cross-hatched, hung from curve c), curves of basinal driving subsidence for $\beta = 2$, 3 & 4 and time-scale of radiometric ages in million years B.P. See text for discussion.

(Llandeilo–Ashgill) and 23% (Pridoli); i.e. average vertical extensions respectively of 45% and 30%. For the Llandeilo–Pridoli this results in a bulk mean shortening of 28% (vertical extension 39%), which is close to that at which slatey cleavage is developed. Since the thick Wenlock–Pridoli is barely or not at all cleaved and the Ordovician is quite strongly cleaved at some levels, this overall estimate seems reasonable. For the Lower Devonian, non-cleaved in most areas, a horizontal shortening of 20% (vertical extension 25%) has been used.

The Arenig–Llanvirn, like the Llandeilo–Lower Devonian, is assumed to have been shortened by about 30% overall, but was also extended during the initial rifting phase associated with basin formation prior to thermal subsidence. This thinning, by a factor $1/\beta$ (Dewey 1982:395), has been compensated using an initial Arenig β of 2.5 declining to 1 at end-Llanvirn. It is for this reason that the Arenig–Llanvirn portion of the thickness/time curve b has greater thickness value than in the non-corrected curve a in Fig. 3.

The tectonically corrected curve b is an underestimate of the true burial history since it is based on thickness data which are not decompacted; the discrepancy being largest at end-Arenig and zero (where extrapolated) in the Middle Devonian. Strictly speaking the backstripping operation should be based on a decompacted burial graph, but this has not been done for two reasons. Firstly, a sensitivity analysis of a worst case (treating the entire succession as mudstone) shows that the ensuing estimate of β when neglecting this effect is only too low by 10–15% and this is probably within the precision of the ages and palaeo-waterdepths adopted in Fig. 3. Secondly, many estimates of tectonic shortening are too low since they neglect the effect of compaction on originally spheroidal objects used as strain markers (Dunnet & Siddans 1971). This effect implies that the strains adopted in correcting curve a to curve b in Fig. 3 are conservative, and since the effects of this would increase the estimate of β, the two effects possibly attributable to compaction are deemed to cancel each other. The numerical operation to produce the backstripped curve c from b utilizes compaction data from Dewey (1982: Fig. 26) with minor adjustments for thick volcanic piles.

Fig. 3 illustrates portions of driving subsidence curves for $\beta = 2$, 3 and 4 in the case that initial crustal thickness was 31.2 km and total lithospheric thickness was 125 km. These values were chosen by Dewey (1982:386) as a reference case in which crustal elevation prior to rifting lies at sea level and total lithospheric thickness is not abnormal. In the S Wales basin this model seems appropriate (see Fig. 2). The crust in southern Britain consolidated in late Proterozoic time at least 300 Ma before basin formation (Hampton & Taylor 1983) and might not be expected to be abnormally thin or thick in the early Ordovician, unless a considerable change in plate-tectonic regime is invoked during the Cambrian.

In deep-water basins the time/depth curve describing basinal driving subsidence falls below that describing the total subsidence after correction for its load induced component by an amount equivalent to palaeo-waterdepth (Dewey 1982:396–397). Thus curve c in Fig. 3 represents a minimum β value unless palaeo-waterdepth estimates can be hung from it. Key palaeo-waterdepth estimates are derived from the Arenig–Llanvirn volcanics (Jones 1969; Lowman & Bloxam 1981:66), the Ashgill–Wenlock basin margin sediments (James 1983 and unpublished) and the basal Devonian coastal/alluvial sediments. Since the post-Ashgill estimates lie adjacent to the basin margin, and thus may be somewhat shallower than those applicable to the basin axis, the β value fitted through these points may be slightly underestimated.

The conclusion from this subsidence analysis is that a stretching factor of about 3 appears satisfactorily to fit the data. This now needs discussion and possible refinement against the geological background.

Basin formation

Timing and development

In S Wales the Cambrian is about 1.2 km thick, of shallow water facies with evidence of nearby emergence of granitic/metamorphic basement and displays no volcanics (George 1970). It is overlain with marked angular unconformity by Arenig, initially in shallow-water facies with volcanics (George 1970) but rapidly deepening into the Llanvirn with development of acid/basic volcanic sequences, well exposed on Ramsey Island (Kokelaar et al. 1985) and along the coast to the largely rhyolitic deposits around Strumble Head (Bevins & Roach 1979). Llanvirn volcanism is

also prominent at Builth Wells (Jones & Pugh 1948, 1949).

This succession suggests that basin formation was rapid and that its timing can be well constrained. The geochemistry of the volcanics suggests a transition over much of Wales at this time from a volcanic arc to a marginal basin (Bevins et al. 1984) and that this arc and basin were developed on continental crust (Lowman & Bloxam 1981;49). Marginal basins on continental crust without crustal rupture are not unusual (Tarney & Windley 1981).

The Caradoc–Wenlock history (Fig. 2) is dominated by the clear distinction between the basin (often with axial infill from the south) and the marginal platform to the southeast (with relatively condensed, often calcareous deposits) across a plexus of faults. The first major transgression across what was probably a peripheral thermal bulge took place in the Caradoc (Fig. 4). The transgression was possibly aided, as was a transgression of similar magnitude (Ziegler et al. 1968; Bridges 1975) in the Telychian stage of the Upper Llandovery (Figs. 5 & 6), by eustatic sea-level rise (Leggett 1980). In the Ashgill, glacio-eustatic regression advanced the slope break across the Severn–Tywi fault (Fig. 5) with the result that a non-depositional slope was created (James 1983). In the Wenlock and, particularly, the Ludlow the sharp distinction between basin and marginal platform is progessively lost and the basin fill shallows up into the Pridoli without major slope breaks until the line of the Church Stretton fault
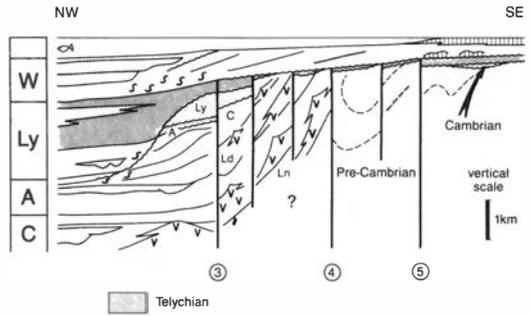

Fig. 5. Sketch section reconstructing Ashgill regressive shelf outbuilding beyond the confines of the basin-margin faults and the late Llandovery (Telychian) transgression across these faults. In this area the earlier Caradoc transgression across the peripheral thermal bulge is not preserved. Basin margin faults schematically represented as vertical and indexed as in Fig. 1. Facies legend as Fig. 2. Length of section approx. 50 km.

(which then formed the basin margin) is reached. Activity on the basin margin faults thus migrated eastwards with time in a general sense and the early rift-defining faults were covered with deposits of the flexural phase of subsidence (cf. Dewey 1982: Fig. 19). The Corris–Llangranog lineament (James & James 1969) in the west of the basin was not active after the Ashgill. It probably lies close to the early bathymetric axis of the basin and below a major vergence divide which formed during later folding (Woodcock 1984b).

During the Pridoli the basin filled to sea level and the succeeding Lower Devonian continues a fluvial infill derived from the NW.

Mechanism and constraints

It has recently been argued (Dewey 1982:386–391) that two contrasting but not mutually exclusive mechanisms for stretching exist; namely the lithospheric attentuation model (M) of McKenzie (1978) and the lithospheric dyking model (R) of Royden et al. (1980). Dewey illustrates these models in evolutionary cross-sections (op. cit., Figs. 17 & 19 respectively) and offers criteria for their distinction. The M model is characterized by listric extensional faulting, weak volcanism, an early peripheral bulge away from the rift shoulder and a β value of less than 2 for the reference crustal and lithospheric thicknesses cited above. The R model has a β value of more than 3 for these same crustal and lithospheric thicknesses and is characterized by steep basin margin faults, prominent rift shoulders and strong volcanism.
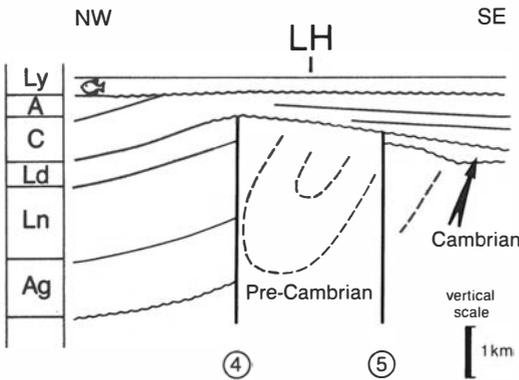

Fig. 4. Sketch section reconstructing Caradoc overstep of the Longmynd Horst, interpreted as a peripheral thermal bulge. Data after Pocock & Whitehead (1948: fig. 19). Basin margin faults schematically represented as vertical and indexed as in Fig. 1. Length of section approx. 27 km.

The key factor in evaluating the basin-forming mechanism in S Wales is the intense early volcanism, and the lack of significant late volcanism. Also significant are the steep basin-marginal faults and the associated hypabyssal intrusions, for example in the Builth Wells–Llandrindod area (Jones & Pugh 1948). There is a strong suggestion at Builth Wells that early uplift of volcanics near a rift-shoulder thermal bulge could explain the classic terraced shoreline phenomena in the New-mead sediments of the Llanvirn described by Jones & Pugh (1949). All the above arguments strongly favour an R model. This in turn suggests that the β value of about 3 is realistic if the assumption of the reference values of a 125 km lithosphere and 31.2 km crust is correct. The only way to produce a β value of about 2 under these assumptions requires that the entire Llandeilo–Pridoli succession has been doubled in thickness by tectonic compression. This possibility is so far out of line with field evidence that it may be discarded.

The driving subsidence curve could however also be fitted with an β value near 2 if the lithosphere was initially appreciably thinner than 125 km, which is quite probable in an area adjacent to an island arc. The estimate of end-Llanvirn initial subsidence that may be made from Fig. 3 is about 2.2 km. With β equal to 2 this subsidence could have resulted from extension of a lithospheric thickness of about 94 km (Dewey 1982: Fig. 13) with a 31.2 km crust.

Fig. 6. Sketch section reconstructing Llandovery–Wenlock overstep of the peripheral thermal bulge defined by the Pontesford lineament near Llandeilo. Data after Williams (1953). Basin margin faults schematically represented as vertical and indexed as in Fig. 1. Faults I, II and III were re-activated with thrust components during Caledonian folding. Length of section approx. 15 km.

There is no evidence from gravity (Griffiths & Gibb 1965) or from the nature of the volcanic record (Stillman & Francis 1979) that full rupture of a back-arc continental crust occurred. This constrains β to be less than about 4.4, at which initial subsistence reaches the average depth of oceanic ridges and oceanic crust is produced (Dewey 1982:392), assuming the crustal and lithospheric reference thicknesses as used in Fig. 3. Moreover, this lack of evidence for crustal rupture demonstrates the requirements for tectonic corrections, since without these a β value of about 5 is obtained. A further important consideration is the geometry of the basin-marginal fault systems. Representative sections across these are given in Figs. 4, 5 and 6. The exact attitude of these faults is not known but Woodcock (1984a) gives a convincing resumé of the arguments that they are steep and have a polyphase history including some strike slip. The steep basin-margin faults of an R model would provide a natural locus for strike-slip movements. Despite the fact that local reverse reactivation of faults in the cover genetically linked to these basement faults has taken place (Fig. 6, see Williams 1953) there is little evidence that the latter formerly possessed any of the characteristics of listric extensional systems (e.g. Gibbs 1984).

The cause of the rapid development of the back-arc basin cannot be approached quantitatively by subsidence analysis. It may well have been linked to the switch from a compressional Chilean-type volcanic arc to an extensional Mariana-type arc (Uyeda 1982). If so, it is perhaps unlikely that the crust at least would be abnormally stretched prior to basin formation. If the change in arc type is caused by along-strike plate buckling during subduction (Bayley) 1982), the lateral extent of the Welsh basin outside its current area of occurrence may not have been large.

In the Lower Devonian the simple back-arc model using an R mechanism breaks down. It is evident from Fig. 3 that thermal subsidence in continuation of that established in the Ordovician and Silurian is not adequate to explain the thickness of the Lower Devonian in the areas immediately marginal to the basin. Moreover, even thicker Devonian sequences than plotted in Fig. 3 are found in Pembrokeshire on the basinal side (north) of the Ritec fault (Dunne 1983). It is likely that the increased subsidence is caused by loading to the NW associated with thickening due to end-Silurian plate collision across the Iapetus suture
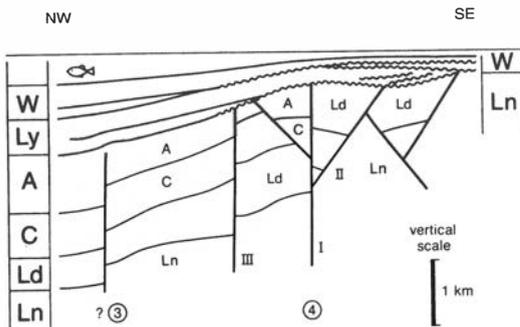
(Phillips et al. 1976). This supports quantitatively a suggestion put forward by Mitchell (1984).

Basin inversion

Timing and development

In S Wales the continuity of sedimentation into the Lower Devonian and the local presence therein of a common cleavage with Silurian and older rocks (e.g. Cope 1979) indicates Middle Devonian as the earliest possible timing for the principal compressional deformation of at least the southeastern portion of the basin. In the Brecon Beacons, Upper Devonian is non-cleaved and lies unconformably above Lower (Turnbridge 1986). Basin inversion probably began in the late Silurian in N Wales, peaked in mid-Devonian across the entire back-arc basin and was complete before the development of the regional base-Carboniferous unconformity north of the Church Stretton fault. The cleavage in Wales, as in much of the southern British Caledonides, shows evidence of formation in a transpressive regime. On a very large scale this regime was probably sinistral (Soper 1986), at least by Devonian times, although the McKenzie & Jackson (1986) model of distributed deformation across an inter-plate contact might yet be able to reconcile local sinistral transpression (e.g. Murphy 1985) with earlier ideas of large-scale dextral regimes along plate boundaries with local irregularities (see Sanderson et al. 1980). The deformation generated both folding and cleavage and the folds now expose a variety of strata having suffered both variable depths of maximum burial and variable strain, both of which, together with temperature, affect metamorphic grade (Roberts & Merriman 1985). The grade is generally low, locally reaching up to greenschist conditions (Bevins & Rowbotham 1983).

Mechanism and constraints

The extreme metamorphic geothermal gradients immediately prior to inversion postulated by Aldridge (1986) are based on underestimates of the thickness of the basin fill, nevertheless the gradients are still very high (40–50°C/km) using the thickness data of Fig. 3. The explanation is probably massive heat convection by dewatering during cleavage formation and it would be unwise to use apparently extreme gradients as a constraint on the possible mechanism of basin inversion. The suggestion by Woodcock (1984b:327) that cleavage might have been generated at depth while upper portions of the basin fill were still accumulating is difficult to reconcile with the simple R model evolution but could apply from Lower Devonian times when a compressive regime was established NW of Wales and stress diffusion to the southeast began.

An attempt at quantitative analysis of the inversion is important and can be constrained by the present-day crustal thickness. This is probably about 35 km from the evidence of deep reflection seismic profiles such as line SWAT 2 shot by the British Institutions Reflection Profiling Syndicate (BIRPS). Moreover, Bouger anomalies suggest no anomalous thickening (Griffiths & Gibb 1965). The crust now comprises up to about 13 km of tectonically thickened basin fill (where subsequent erosion is relatively small) and a pre-basin basement originally thinned during basin formation and later tectonically thickened, in whole or in part, dependent upon the possibility of decollement (Coward & Siddans 1979). The following calculations assume a 31.2 km crust as discussed and used above. For $\beta = 3$, stretching of this crust to 10.4 km and addition of 10.5 km of basin fill would require overall vertical extension of 67% to reach 35 km. This would imply a basement thickening greatly in excess of the probable 35% vertical extension of the post-Llanvirn basin fill. Such differential thickening seems unlikely. With unusually thin original lithosphere (94 km, see above), such that an R-type mechanism could operate at $\beta = 2$, stretching of the crust to 15.6 km and addition of 10.5 km of basin fill would require overall vertical extension of 34% to reach 35 km. This would imply basement and basin fill deforming about equally. Mechanically, this appears quite plausible and would probably be accommodated in the basement by the back-stacking of old normal faults. If crust and lithosphere were both originally unusually thin, a β of about 3 would again be possible. With a crust of 24 km, stretching to 8 km and addition of 10.5 km of basin fill would require a subsequent overall vertical tectonic extension of 89% to reach 35 km. This is not consistent with field evidence in S Wales, but could be more applicable in N Wales where compressional deformation is locally rather more intense (Rast 1969).

Conclusion

The development of the basin may be summarized as follows:

(a) Stretching associated with lithospheric dyking in the Arenig–Llanvirn with prominent rift margins due to thermal uplift. This stretching can be modelled with $\beta = 3$ for 'normal' crustal and lithospheric thicknesses but the strong evidence for an R model can also be accommodated with $\beta = 2$ for 'normal' crust and an attenuated lithosphere of about 94 km thickness. This latter alternative has the merit that it is both plausible in a near-arc environment and necessitates a vertical tectonic extension to likely present-day crustal thickness, which is consistent with the observed later deformation.

(b) Thermal subsidence and sediment loading from Llandeilo to Pridoli with first overstep of the peripheral thermal bulge in the Caradoc.

(c) Renewed subsidence by load-induced flexure in Lower Devonian times related to collisional orogeny to the NW followed by Middle Devonian uplift and cleavage formation related to the southeastwards diffusion of stress associated with this orogeny.

The success of the simple R model presented in this paper makes it unneccessary to postulate that the S Wales basin is of transtensile origin. Although some pre-Llandovery strike-slip undoubtedly took place its effect on sedimentation and basin geometry was probably minor.

The analysis of basin formation by subsidence history is a valuable tool but is intrisically constrained by lack of data on original crustal and lithospheric thicknesses. One way of constraining the various possible values for these parameters is to consider the inversion history and the development of the present-day value for the crust.

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