Chapter 2 Geological framework

The regional setting — Variscan Orogen

A large part of the geology of Europe from Poland to Iberia is represented by the broad, sinuous, roughly E–W-trending Variscan fold belt which has affinities with the Appalachian Belt on the other side of the Atlantic Ocean and the Uralide Belt in Russia. This section provides a very generalized (and over-simplified) picture of the European Variscides within which to locate the particular magmatic rocks from south-west England which form the substance of this volume. Detailed reviews of the tectonic framework and development of the Variscan fold belt and its different segments may be found in Krebs (1977), Zwart and Dornsiepen (1978), Franke and Engel (1982), Behr *et al.* (1984), Lorenz and Nicholls (1984), Matte (1986), Ziegler (1986) and Franke (1989). The following outline largely follows Franke (1989).

The Variscan Orogen of northern Europe consists of a series of Ordovician to Carboniferous rift-generated basins, separated by metamorphosed crystalline ridges, which were progressively closed by the northward migration and subsequent collision of African Gondwanaland with northern Baltica. Early work by Kossmat (1927) subdivided the Variscan orogenic belt into a number of tectonic zones separated by major thrusts, ranging from the northern (external) Rhenoher-cynian zone to the (internal) Saxothuringian and Moldanubian zones of central Europe (Figure 2.1). The Rhenohercynian and Saxothuringian zones include basinal sedimentary sequences and volcanics indicative of initial rifting and later synorogenic sandstones heralding subsequent closure. Although much of the basic–acid volcanism of these basins is typical of rifted continental crust, basaltic lavas with mid-ocean ridge (MORB) chemical characteristics are present and suggest that narrow oceans floored by oceanic crust were generated. Between these two zones in central Europe is the Mid-German Crystalline Rise largely composed of pre-Devonian magmatic and high-grade metamorphic rocks that have their counterpart in Armorica as the crystalline Normannia High. The Moldanubian zone is dominated by largely Precambrian magmatic and metamorphic rocks overprinted by Variscan tectonometamorphic events.

Essential to the Variscan story is the staged closure of the sedimentary basins by subduction of narrow oceanic segments and northwards-directed thrusting from late Ordovician through to the Carboniferous, such that many sequences are thrust bound and parautochthonous or allochthonous in character. In contrast to the general northwards convergence and progressive closure throughout much of the Palaeozoic, the basinal region that formed in the Rhenohercynian zone was not only late in development, but rapidly closed – opening in early Devonian and closing by the end of the late Devonian.

The magmatic rocks described in this volume are all within the external Rhenohercynian Zone of the Variscides, some of which have features in common with their counterparts in the rest of the orogen. The sedimentary and volcanic record of the Rhenohercynian Zone reflects an early Devonian rifting event that was rapidly followed by late Devonian to late Carboniferous crustal shortening that produced stacks of northwards-converging nappes and accompanying low-grade regional metamorphism. Relatively high-level, post-orogenic, granite magmatism is a common feature of this zone. There is a difference of opinion on the plate-tectonic situation of the initial rifting episode in the Rhenohercynian zone, but it is generally agreed that crustal attenuation was sufficient for the production of some oceanic crust. Four models have been suggested:

1. back-arc basin related to northerly directed subduction of an oceanic area to the south of Armorica (Floyd, 1982b; Leeder, 1982; Ziegler, 1986);
2. intracratonic strike-slip pull-apart basin (Bad-ham, 1982; Barnes and Andrews, 1986);
3. small ocean basin related to southerly directed and although the older intrusions originated by subduction under an active arc to the north of Armorica (Holder and Leveridge, 1986);
4. small ocean basin generated as continental crust overrode a relict Caledonian spreading axis (Matthews, 1978; Franke, 1989).

Basic–acidic volcanism is a characteristic of all Variscan tectonic zones, although volcanic sequences are generally dominated by basalts ('spilites') with minor trachytes and rhyolites ('keratophyres' and 'quartz keratophyres'). Early

Devonian rhyolites and late Devonian–Dinantian basalt pillow lavas and intrusives in south-west England have their temporal counterparts throughout the Rhenohercynian and Saxothuring-ian Zones, although the Moldanubian zone also exhibits late Dinantian calc-alkaline andesite-dominated volcanism. Throughout the Rhenohercynian zone most of the basalts are incompatible element-enriched, intraplate tholeiites and alkali basalts – the latter of which are particularly characteristic of south-west England (Floyd, 1982a; Wedepohl *et at,* 1983). However, Middle and Upper Devonian basalts within stratigraphically restricted *mélange* sequences have chemical affinities to MORB (Floyd, 1984; Grosser and Dorr, 1986) and, together with the Lizard ophiolite and Start Complex in Cornubia, they provide evidence for the existence of Rhenohercynian oceanic crust. Another chemical characteristic that appears to be a common feature of the basalts within the Rhenohercynian (and possibly Saxothuringian) Zones is the change in degree of incompatible-element enrichment from the Devonian to the Carboniferous (Floyd, 1982a). This is probably a reflection of the change from generally depleted MORB-type basalts in mid-late Devonian rift zones to the more enriched intraplate basalts of late Devonian and early Carboniferous age on the margins of basins.

As far as the Variscan granites are concerned, all the tectonic zones contain granitic massifs, many of which include types resembling those of the Cornubian batholith. In addition to older pre-Variscan intrusives and metamorphics, two or three distinct intrusive phases are found, as illustrated below.

The Kru■né hory Mountains granitoids (northwest part of the Bohemian Massif) in the Saxothuringian Zone, have ages ranging from 340 to 260 Ma (K/Ar method). They are grouped into an Older Intrusive Complex, with biotite and I-type granite characteristics, and a Younger Intrusive Complex, with Li-mica and the chemical attributes of S-type granites. The youngest rocks also have high concentrations of Rb (c 500 ppm), Sn (*c.* 38 ppm) and F (*c.* 2900 ppm) (temprok, 1986). The granites of Normandy and northern Brittany group into suites with ages ranging from 340–300 Ma and 300–280 Ma (both K/Ar and Rb/Sr methods). Where these have been characterized as 'S' or 'I' types, the latter tend to be the older, and the younger group tends to be relatively enriched in Sn and other metals (Chauris *et al.,* 1969; Adams, 1976; Peucat *et al.,* 1984; Georget *et al.,* 1986).

In the Schwarzwald, granites (dated by the Rb/Sr method) form three series ranging from about 363 Ma through 330–310 Ma (the majority) to 290–280 Ma. They lie in the Moldanubian zone and although the older intrusions originated by crustal melting, the younger show evidence of fractional crystallization and magmatic differentiation, together with high-level intrusive characteristics. With decreasing age, overall Sr contents fall (245 to ?24 ppm), whereas both Rb and F rise (228 to 483 ppm Rb and 618 to 1575 ppm F) (Emmerman, 1977). Also in the Moldanubian zone are the granitoids of southern Brittany and the Massif Central and here, too, there are 'older' and 'younger' generations dated at about 310–300 Ma and 290–280 Ma (K/Ar and Rb/Sr methods) (de Albuquerque, 1971; Reavy, 1989). In both cases there is evidence of differentiation sequences, but whereas de Albuquerque (1971) supposes that the parent rocks were pelites and greywackes, Pinto (1983) suggests that the younger granites have I-type characteristics.

Throughout the Variscides of northern Europe, therefore, granites of approximately the same age as those in Cornubia are found and they often display similar chemical characters, such as being enriched in Li, Rb, Sn and F. There are two important differences, however. On the continental mainland, they lie overwhelmingly within the Saxothuringian and Moldanubian tectonic zones, whereas the Cornubian rocks are in the Rhenohercynian zone where rather few granitoids (for example, Harz Massif) occur relative to the mainland. Secondly, in Cornubia, only the younger (300–280 Ma) suite is found, the older generations apparently being absent. These features tend to support the suggestion, discussed elsewhere in this volume, that the Cornubian batho-lith may be detached from its original roots.

The local setting – Rhenohercynian zone of south-west England

We now turn to south-west England, which is representative of the western end of the external Rhenohercynian zone. This section provides a simplified geological framework for the volcanic and plutonic events outlined in Chapter 1, and from which the sites have been chosen. Recent work and summaries of the geology of south-west England may be found in Durrance and Laming (1982), Hancock (1983), Hutton and Sanderson (1984) and Dineley (1986). An outline geology of south-west England and the distribution of magmatic rocks is shown in (Figure 1.1).

Throughout the Devonian and Carboniferous periods, south-west England was mainly characterized by the successive northward development and subsequent closure of small, often deep-water, basins (Matthews, 1977). They were bordered to the north by shallow coastal shelves and initially received debris from the Old Red Sandstone continent in South Wales and the Bristol Channel (Freshney *et al.,* 1972, 1979; Tunbridge, 1986). Later, a southern source was also important, with flysch and sedimentary *mélange* material derived from northwards-advancing nappes filling both late Devonian and Carboniferous basins (Isaac *et al.,* 1982; Holder and Leveridge, 1986). During mid-Devonian times, basins in the east were segmented by 'rises' that developed shallow-water carbonate banks possibly capping volcanic seamounts (Edmonds *et al.,* 1969).

Structurally, the Variscan fold belt is dominated by thrust ('thin-skin') tectonics with piles of southerly derived nappes characterizing Cornwall and central south-west England (for example, Matthews, 1981; Isaac *et al.,* 1982; Shackleton *et al.,* 1982; Leveridge *et al.,* 1984; Selwood and Thomas, 1986a and 1986b). (Figure 2.2) (from Dineley, 1986) summarizes the main structural features of Variscan South-west England. From a broad correlation of K/Ar age dates on slates and phyllites (Dodson and Rex, 1971) with structural zones (Dearman *et al.,* 1969; Dearman, 1969, 1971; Sanderson and Dearman, 1973), minimum ages for deformation and related metamorphism can be estimated. In western Cornwall the initial F_1 deformation (370–350 Ma BP) occurred at the end of the Devonian, whereas elsewhere it represents a mid-Carboniferous event (345325 Ma BP). The second deformation phase (F_2) overprinted both areas (Sanderson, 1973) during the late Carboniferous (315–275 Ma BP). Two subsequent, relatively minor, fold phases ($F_{3–4}$) are probably related to the emplacement of the batholith.

Volcanic sequences occur at various structural levels within the nappes such that Upper Devonian lavas can be found overlying Lower Carboniferous volcanics of similar eruptive character. For example, detailed structural schemes for central south-west England have been recorded by Isaac *et al.* (1982), Selwood *et al.* (1985) and Selwood and Thomas (1986a, 1986b), and these emphasize the presence of various nappes thrust over parautochthonous/autochthonous 'basement'. (Figure 2.3) summarizes the relationship between recognized structural units in this area: the majority of the Upper Palaeozoic volcanics occur within the Port Isaac and Tredorn Nappes (north Cornwall), the Greystone Nappe (central area between Bodmin and Dartmoor) and the Chudleigh Nappe (east Dartmoor). The Greystone Nappe is considered to be an early, synsedimentary, gravity-transported slice emplaced in the late Viséan and subsequently overidden by other 'younger' nappes (for example, Port Isaac Nappe, with mid-Carboniferous ages; Tredorn Nappe, with late Carboniferous ages) derived from further south (Isaac *et al.,* 1982; Selwood and Thomas, 1986b). Basically, tectonic instability was heralded not only by prograding flysch and sedimentary *mélanges,* but also by volcanic activity concentrated within the late Devonian–Dinantian period, both coeval with, and prior to, the advancing deformation front.

We shall now consider in more detail the stratigraphical position and environmental significance of the various magmatic groups outlined in Chapter 1 and how they relate to the geological development of south-west England.

Devonian volcanic activity

Lizard and west Cornwall region

The most significant part of the magmatic story starts in southern Cornwall with the development of the Gramscatho Basin in the late, early Devonian (Barnes and Andrews, 1986). The Gramscatho Group ((Figure 2.4) and (Figure 2.5)) comprises a deep-water turbiditic sandstone–mudstone association and can be divided into two main parts (Holder and Leveridge, 1986):

1. a northern parautochthonous region which is separated by the Carrick Thrust (Leveridge *et al.,* 1984) from
2. an allochthon composed of a series of thrust slices which includes the Lizard Complex at the highest structural level.

The Gramscatho Group is generally considered to be mid- to late Devonian in age (Sadler, 1973; Le Gall *et al.,* 1985); it is tectonically juxtaposed against the early Devonian shallow-water Mead-foot Group along the Perranporth–Pentewan Line (Dearman, 1971).

The importance of the magmatic rocks in this region (Figure 2.2) centres on the interpretation of the Lizard ophiolite as the oceanic basement to the Gramscatho Basin (Barnes and Andrews, 1986; Holder and Leveridge, 1986), and the close association of various, largely mid–late Devonian, basaltic volcanics that have chemical compositions akin to mid-ocean ridge basalts (MORB) (Floyd, 1982a, 1984; Barnes, 1984). The ophiolite and exotic volcanics thus constitute the only good evidence for oceanic crust in this part of the Rhenohercynian belt. The following magmatic suites can be identified:

1. Lizard ophiolite (Strong *et al.,* 1975; Bromley, 1976, 1979; Styles and Kirby, 1980; Kirby, 1979a, 1979b, 1984) and associated hornblende schists.
2. Metabasic clasts within the south Cornish *mélange* zone (Barnes, 1983) – the Roseland Breccia Formation of Holder and Leveridge (1986) ((Figure 2.4) and (Figure 2.5) – and possibly including the pillow lavas of Mullion Island.
3. *In situ* Middle Devonian submarine volcanics (Tubbs Mill unit) at the base of an allochthonous slice of the Gramscatho Group.
4. Basic, submarine volcanics and associated high-level intrusives within the late Devonian Mylor Slate Formation of west Cornwall (Taylor and Wilson, 1975).
5. An additional basic sequence (undifferentiated Devonian age, possibly early Devonian; Dine-ley, 1986) are the metavolcanic greenschists of the Start region. Magmatically and tectonically, they could form part of the south Cornish allochthonous terrane and represent another segment of oceanic crust thrust over the early Devonian sediments to the north (Coward and McClay, 1983). There is evidence to suggest that they may well form part of a more extensive 'metamorphic basement' to the sea floor to the south and west of Start Point (Phillips, 1964).

A simplified interpretation of events during the mid- to late Devonian in this region might be as follows. The Gramscatho Basin was generated initially by extension and thinning of continental crust (Matthews, 1977; Leeder, 1982). The attendant magmatism eventually developed oceanic crust that subsequently formed the base to part of the basin (Rattey and Sanderson, 1984; Barnes and Andrews, 1986) and is now represented by the Lizard ophiolite and exotic volcanics within the south Cornish *mélange* (groups 1 and 2 above). The latter have a similar chemistry to the Tubbs Mill volcanics (group 3), which appear to be the earliest (middle Devonian), contemporaneous, submarine lavas now located at the thrust-terminated base of one of the flysch units. The basin was filled from the south by turbiditic sandstones (Gramscatho Group) derived from the erosion of nappes predominantly composed of acidic continental arc materials (Floyd and Leveridge, 1987; Shail and Floyd, 1988). The deepening of the basin is marked by the parautochthonous argillites of the late Devonian Mylor Slate Formation which are probably coeval with the upper part of the northwards-advancing Grams-catho turbidites and *mélanges* (Holder and Leveridge, 1986). Within the Mylor Slate Formation are discrete pillow lava and intrusive horizons (group 4 above) which are chemically distinct from the earlier MORB-like volcanics mentioned above. The filling of the basin is indicated by a proximal heterogeneous sedimentary *mélange* composed of MORB-like volcanic clasts and continental-derived rudaceous debris (Barnes, 1984). Suboceanic slicing and metamorphism of the ocean crust (Vearncombe, 1980), soon after its formation, heralded the onset of obduction, which eventually emplaced the Lizard ophiolite on top of a stack of Gramscatho nappes and the *mélange* towards the end of the Devonian (Styles and Rundle, 1984).

Central south-west England region

To the east of the Perranporth–Pentewan Line, which marks the boundary of the Lizard/west Cornwall region, is the magmatically distinct central region, that for the convenience of this summary covers north Cornwall and Devon. Idealized stratigraphical relationships in Devon are illustrated in (Figure 2.6).

The early Devonian (Dartmouth Slate, Staddon Grit and Meadfoot Formations) in south Devon is characterized by shallow-water tidal facies as well as non-marine fluvial and alluvial plain sequences (Dineley, 1966; Richter, 1967; Selwood and Durrance, 1982; Pound, 1983). The earliest record of Devonian volcanism in south-west England is seen within the Dartmouth Slates; this comprises limited rhyolitic and basaltic lavas and intrusives with spatially associated tuffs (Shannon and Annis, 1933; Durrance, 1985a). The rhyolites are typical of an 'active margin environment' (Durrance, 1985a) and may have been produced by the melting of the underlying continental crust. During the middle Devonian, more open marine, shallow-shelf conditions prevailed with the local development of massive carbonate platforms (for example, Plymouth Limestone) which formed an east–west ridge extending from Dartmouth and Plymouth westwards towards Bodmin (Selwood and Durrance, 1982; Burton and Tanner, 1986). Volcanism was again restricted to relatively small volume, but numerous, tuff horizons and minor basaltic, highly vesicular lavas. Acid and basic volcaniclastics (for example, Kingsteignton Volcanics) are best developed in south-east Devon (Tomes, Newton Abbot), where they interdigitate with reef limestones (Ussher, 1912; Middleton, 1960; Richter, 1965). A completely different and interesting igneous association is the Clicker Tor ultramafic body and associated basic lavas and volcaniclastics near Liskeard (Burton and Tanner, 1986) developed within middle Devonian argillites and tuffaceous slates.

In a similar fashion, initially shallow-water sedimentation in north Devon started to the south of the broad Old Red Sandstone coastal plain, and throughout the Devonian was characterized by a migratory sequence of interdigitating marine and non-marine elastic sediments (Webby, 1966; Tunbridge, 1983). No significant magmatic activity is associated with this regime, in contrast to south Devon over the same time period.

From the magmatic viewpoint, significant and widespread volcanic activity did not start until the late Devonian and its products now form a roughly linear belt from the north Cornish coast (Padstow) to east of Dartmoor at Chipley (Figure 1.1). The generally shallow-water conditions over the whole area during the early Devonian were terminated by differential subsidence at the southern margin of the shelf; with the development of a basinal facies during the mid and late Devonian. A schematic diagram showing the general relationship between the various facies during the Devonian and Carboniferous is illustrated in (Figure 2.7).

The northern shelf–basin margin is interpreted as a half-graben fracture system (Selwood and Thomas, 1986b) and has a linear belt of volcanic activity closely associated with it. Extensive volcanism thus appears to be associated with the development of the basin, but related to marginal fractures rather than forming an incipient basement. This is probably a similar situation to the submarine lavas within the Mylor Slate Formation of west Cornwall of comparable age and clearly distinct from the Lizard and the *mélange* volcanics. Classic pillow lava sequences (for example, at Pentire Point and Chipley) are seen within the deep-water argillites of the late Devonian basinal facies (Dewey, 1914; Gauss and House, 1972, Middleton, 1960). High-level doleritic intrusives are also common in the Padstow area, and exhibit two petrographically and chemically distinct primary suites (Floyd, 1976; Floyd and Row-botham, 1982). Rare ultramafic material, found within late Devonian slates, also occurs within the volcanic belt and, spatially, this is closely associated with dolerites and basalts. Such a body is the Polyphant Complex to the north-east of Bodmin Moor (Reid *et al.,* 1911; Chandler *et al.,* 1984), which in the tectonic scheme of Isaac *et al.* (1982) is considered to be part of an ophiolitic *mélange* possibly equivalent to the Gramscatho Basin *mélanges* (Isaac, 1985).

To summarize, a number of magmatic suites can be identified that are broadly related to their eruptive setting and age:

1. Minor early Devonian rhyolites, basic lavas and volcaniclastics that are representative of the earliest expressions of magmatism in the Variscan fold belt (south Devon).
2. Volcaniclastic-dominated sequences and shallow-water basaltic lavas within the middle Devonian, that are often associated with the margins of carbonate reefs and platforms (south and south-east Devon).
3. Deep-water pillow lavas and high-level sills related to the northern, fractured margin of the late Devonian Trevone Basin (north Cornwall).
4. Rare association of ultramafic intrusives with late Devonian basic volcanics and deep-sea sediments (east Bodmin Moor).

Carboniferous volcanic activity

Sedimentation in the early Carboniferous consisted, in north Cornwall and north Devon, of deep-water argillites deposited in basins that spread northwards across the shelf margin (Goldring, 1962; Matthews, 1977). However, in south-east Cornwall and south Devon, a shallow-water paralic facies dominated, and the Devonian reefs and carbonate platforms became lagoonal areas (Whiteley, 1984; Selwood and Thomas, 1986b). Due to a rising source region to the south, prograding flysch deposits began to spread northwards across the whole area, eventually filling and shallowing the northern basins by the Namurian (Selwood and Thomas, 1986b). Subsequent sedimentary history in the late Carboniferous reflects the continuation of basin filling by clastic materials until largely non-marine condi tions prevailed (Freshney and Taylor, 1972; Higgs, 1984). Magmatic activity was, however, mainly confined to the early Carboniferous in the north of the region, and terminated prior to the onset of the Namurian flysch phase.

Carboniferous volcanicity was thus temporally restricted (Viséan–Tournaisian) and again concentrated along shelf–basin margins. It forms a broad belt from the north Cornish coast (Tintagel) inland to central south-west England and around Dartmoor to the Teign Valley (Figure 1.1). The Carboniferous belt roughly parallels late Devonian activity, but lies further north, reflecting the northward migration of the Carboniferous basins.

On the basis of macrofaunal evidence in the associated sediments, the Tintagel Volcanic Formation is correlated with the Meldon volcanics (north-west Dartmoor), representing volcaniclastic-dominated eruptions dated at near the Viséan–Tournaisian boundary (Francis, 1970). The Tintagel volcanics are enclosed in pyritous black mudstones suggesting a deep-water basinal eruptive setting, whereas the Meldon volcanics are associated with calcareous argillites, thin limestones and cherts that probably flanked a rise slope. Owing to the extensive within-bed shearing and the late development of white mica, the Tintagel volcanics are now heterogeneous greenschists with identifiable basalt lava only preserved as relicts within low-strain lenses. Geochemical data demonstrate that they are a comagmatic series with alkali-basalt affinities (Robinson and Sexton, 1987; Rice-Birchall and Floyd, 1988).

The central south-west England thrust–nappe terrane between Bodmin Moor and Tavistock includes acid and basic volcaniclastics, vesicular (commonly pillowed) basaltic lavas and thick basic intrusives (dolerites and some gabbros) –the latter being typical of the greenstones of Variscan south-west England. Some of the massive basic bodies were intruded at a high level and may show pillowed tops. The majority of these volcanics are late Viséan to late Tournaisian in age (Stewart, 1981) and confined to the early Greystone Nappe. The volcanics are associated with deep-water, black mudstones, distal turbiditic mudstones, radiolarian cherts and olistostromic material, and they were developed in a small pelagic basin over a 19 Ma period (Chandler and Isaac, 1982). The interest in this pelagic–olistostrome–volcanic association concerns its interpretation as an ocean-floor sequence, with the basalts and dolerites apparently exhibiting chemical affinities with oceanic crust (Chandler and Isaac, 1982). On this basis and the (possible) initial close stratigraphical association of the late Devonian mafic/ultramafic Polyphant Complex, it has been suggested that this was an ophiolite association (Chandler and Isaac, 1982) and that the marginal basin model for the Devonian (for example, Leeder, 1982) could be applicable to the early Carboniferous. The evidence on which this interpretation is based is considered tenuous at best (Selwood and Thomas, 1986b) and the application of the geochemical data is open to question. A completely different eruptive setting is exhibited by apparently isolated volcanic centres, such as that seen at Brent Tor (north of Tavistock) where a reworked apron of hyaloclas-tite debris bordering a seamount-like edifice occurs.

To the east of Dartmoor in the Teign Valley, relatively minor pillow lavas and acidic volcanic- lastics are interbedded with radiolarian cherts, black limestones and mudstones, although much of the activity is characterized by huge, massive, basic intrusives (greenstones). Many of these bodies are dolerite sills or nearly concordant sheets (sometimes 100 m or more thick) showing internal differentiation, with the development of granophyric fractions (Morton and Smith, 1971).

In summary, the volcanics have broadly similar eruptive ages and volcanic assemblages, although a number of suites can be recognized by their tectonic association (Figure 2.3):

1. Tintagel Volcanic Formation basaltic volcaniclastics and minor intrusives within the Tredorn Nappe (north Cornish coast and inland).
2. Volcaniclastics, pillow lavas, and massive intru-sives mainly within the Greystone Nappe (between Bodmin Moor and Tavistock).
3. Volcaniclastics, minor lavas and massive intru-sives within the Blackdown Nappe (west Dartmoor).
4. Minor volcaniclastics and pillow lavas, with dominant massive intrusives within the autoch-thon (Teign Valley, north-east Dartmoor).

Carboniferous–Permian plutonic activity

The geography of Cornwall and west Devon is dominated by the granite batholith which constitutes the spine of the peninsula. Like all batholiths, it is made up of many individual intrusions embracing a range of ages and compositions. Although small by global standards and lacking the dioritic and granodioritic components often seen, it compensates for these shortcomings by showing an unrivalled sequence of late- and post-magmatic processes, including mineralization.

The detailed relationship of the batholith with the structural environment outlined above is still somewhat obscure, but from geophysical evidence it is known to extend from a few kilometres east of Dartmoor into the Atlantic west of the Isles of Scilly, a distance of about 250 km. Beyond that, a row of gravity 'lows' extends WSW to the continental margin, and Day and Williams (1970) have interpreted these as being due to granite. If the 'lows' are part of the peninsular granite, then they would extend the length of the batholith to some 500–600 km. Over the Dartmoor–Scilly Isles distance, the batholith varies from about 20 km thick in the east, to half that thickness in the west, and is 40–60 km wide at its base. The Mohorovi■i■ discontinuity beneath southwest England occurs at a depth of approximately 27–30 km (Bott *et al.,* 1958; Bott *et al.,* 1970; Holder and Bott, 1971; Tombs, 1977; Shackleton *et al.,* 1982; Brooks *et al.,* 1983; Brooks *et al.,* 1984). About 95 km WNW of the Isles of Scilly is the small submarine granite outcrop of Haig Fras. This has affinities with the Cornubian batholith and has been interpreted as being separated from it by wrench faulting (Exley, 1966). However, it also contains gneissose material and an alternative suggestion is that it may represent basement that has been thrust northward in a nappe (Goode and Merriman, 1987). A third possibility is that it is related to separate submarine granites of unknown age rumoured to be present closer to Ireland. Day and Williams (1970), for example, consider the Ladabie Bank gravity 'low' (about 40 km north-west of Haig Fras) to be due to granite, but no granite north of Haig Fras was noted in the SWAT seismic survey (BIRPS and ECORS, 1986). Edwards (1984), on the other hand, believes that Haig Fras belongs to a separate batholith parallel to the one exposed on the mainland, and he notes that both have been displaced dextrally.

The most substantial geophysical evidence as to the nature of the base of the batholith comes from Brooks *et al.* (1984) who have found a southwards-dipping seismic reflector at a depth of about 10–15 km. This is the level at which the densities of country rock and granite merge and Brooks *et al.* (1984) interpret this reflector as a thrust. Although this conforms well with the general structural picture, it creates a problem by suggesting a separation of the batholith from its original roots and thus from what would normally be the source of hydrothermal and metallic constituents. Shackleton *et al.* (1982) have included similar thrusts in their interpretation and have proposed that 'the granites were generated farther south and injected northwards as a sheet-like body from which protrusions moved diapirically upwards and by stoping to form the separate exposed granite masses'. They do not specify whether this sheet was magmatic, and their suggestion creates other problems to do with heat, density and vertical and horizontal granite movement, to the extent that granite hot enough to rise diapirically would not also be expected to travel subhorizontally for perhaps 200 km.

The upper surface of the batholith is exposed in a series of seven major outcrops. These are (Figure 1.1), from east to west: the Dartmoor, Bodmin Moor, St Austell, Carnmenellis, Tregonning–Godolphin, Land's End and Isles of Scilly bodies (approximate exposed areas are, respectively, 650 km^2, 220 km^2, 85 km^2, 135 km^2, 13 km^2, 190 km^2 and 30 km^2). Several minor bodies also crop out and include Hemerdon Ball (south of Dartmoor), Hingston Down and Kit Hill (between Dartmoor and Bodmin Moor), Belowda Beacon and Castle-an-Dinas (north of St Austell), Cligga Head, St Agnes Beacon, Carn Brea and Carn Marth (north of Carnmenellis) and St Michael's Mount in Mounts Bay, east of the Land's End Peninsula (Figure 1.1). Each of the major outcrops is itself made up of more than one intrusion, and recent work (Hill and Manning, 1987) suggests that some may contain many more intrusions than hitherto suspected.

The ages of the exposed biotite-bearing granites indicate that the main period of magmatism lay between 290 Ma and 280 Ma BP, and that this was followed by a second phase some 10 Ma later (Darbyshire and Shepherd, 1985; (Table 2.1)). The source of the early granites has always been regarded as crustal, largely because of the nature of their included xenolithic material, and this view has been reinforced in recent years by determinations of their initial $^{87}Sr/^{86}Sr$ ratios (Table 2.1), Pb and Sr isotopes (Hampton and Taylor, 1983), and general chemistry and mineralogy which show such features as high K_2O/Na_2O and low Fe_2O_3/FeO ratios, high normative corundum, and rare or absent hornblende, sphene and magnetite. These are all characteristics of the 'S-type' granites *sensu* Chappell and White (1974). Most authors suggest partial melting of some such lower-crustal source rock as poorly hydrated garnet granulite (Charoy, 1979; Floyd *et al.,* 1983; Stone and Exley, 1986; Stone, 1988) or cordierite–sillimanite–spinel gneiss 'similar to Brioverian basement' (Charoy, 1986), a suggestion supported by the presence of sillimanite-bearing pelitic xenoliths within the Bodmin Moor and Carnmenellis Granites (Ghosh, 1927; Jefferies, 1985a, 1988). The predominantly pelitic character of the source is emphasized by the high $\delta^{18}O$ content of the granite (10.8–13.2‰; Sheppard, 1977; Jackson *et al.,* 1982) and their high average ammonium content (36 ppm; Hall, 1988).

The partial melting of the crustal component is envisaged as occurring at pressures of 7–8 kbar and temperatures of about 800°C and, by the time that it reached the present level of erosion and exposure, the granite was manifestly hydrated.

(Table 2.1) Ages and initial Sr isotopic ratios of granitic rocks from the Cornubian batholith (data from Darbyshire and Shepherd, 1985, 1987)

Intrusive phase	Outcrop and granite type	Rb-Sr age (Ma)	Initial 78 sr/86sr ratio	Comments
Major	Dartmoor (B)	280 ± 1	0.7101 ± 0.0004	—
	Bodmin Moor (B)	287 ± 2	0.7140 ± 0.0002	Mineral age
	St Austell (B)	285 ± 4	0.7095 ± 0.0009	—
	Carnm.enellis (B)	290 ± 2	0.7130 ± 0.0020	Mineral age
	Tregonning (E)	280 ± 4	0.71498 ± 0.00381	Highly evolved, lithium-rich
	Land's End (B)	268 ± 2	0.7133 ± 0.0006	Mineralization re-set age
Minor	Hemerdon Ball	304 ± 23	0.70719 ± 0.01025	Heavily mineralized
	Kit Hill	290 ± 7	0.70936 ± 0.00228	—
	Hingston Down	282 ± 8	0.71050 ± 0.00119	—
	Castle-an-Dinas	270 ± 2	0.71358 ± 0.00122	Later intrusion re-set age
	Carn Marth	298 ± 6	0.70693 ± 0.00207	—
Dykes	Meldon 'Aplite'	279 ± 2	0.7098 ± 0.0017	
	Brannel Elvan	270 ± 9	0.7149 ± 0.0031	Re-analysed
	Wherry Elvan	282 ± 6	0.7120 ± 0.0025	Re-analysed
Mineral veins	South Crofty	269 ± 4		—
	Geevor	270 ± 15	0.7122 ± 0.0012	—

It is believed that this state was achieved by the drawing-in of water from the surrounding rocks. That the granite was also cool is evident from the low grade of thermal metamorphism displayed by the enveloping sediments.

In addition to the crustal component, there is an enrichment in such elements as U, Th, F, Cl and Sn, Cu, W, not all of which could easily have been contributed by incorporation of normal crustal material, and which must, therefore, have been derived either from a previously enriched source or by a direct addition from the mantle. The general conclusion (Simpson *et al.,* 1976; Simpson *et al.,* 1979; Lister, 1984; Watson *et al.,* 1984; Thorpe *et al.,* 1986; Thorpe, 1987) is that the latter is more likely, with an addition of radiogenic, halogen and some metallic elements from the mantle. Jefferies (1984, 1988), in a study of radioactive minerals from the Carnmenellis Granite, concluded that these were of magmatic origin, their radio-element content having been derived from fluids circulating in the upper mantle.

Apart from the possible restite origin of some of the xenoliths (Jefferies, 1985a, 1988), the great majority are high-level metapelites or semipelites. Bromley and Holl (1986) have calculated the settling rates of the xenoliths. Most would have been enveloped by the stoping action of the rising magma and sunk within the magma body. They suggest, from their findings, that a seismic reflector found within the batholith at a depth of about 5–6 km. (Brooks *et al.,* 1983) might mark the top of the 'zone of xenolith accumulation'.

(Table 2.2) Main evolution and alteration stages of the St Austell Granite (after Bristow et *al.,* in press)

Process	Age (millions of years) *	Depth (km)	f Stage Temperature (°C)	Salinity of fluids	Source of heat	Direction of least stress	Main changes in mineralogy			Associated metalliferous mineralization	Comments
							Feldspar	Quartz	Mica		

I	Emplacement of biotite granite, forming main batholith	290–285	? 3	500–600		Magmatic	Variscan (E–W)	—	—	—	—	Biotite granite which now forms eastern part of the St Austell granite
II	First phase of post-magmatic alteration and mineralization	285–275	2–3	500–?200	Moderate	Magmatic	Initially E–W, then N–S	Limited greisenization alongside veins	—	—	Sn, W	Early greisenization and mineralization e.g. Castle-an-Dinas (W)
IIIa	Emplacement of evolved lithium-rich granites and biotite granites in western part of St Austell granite	275–270	2–3	500–600	—	Magmatic	N–S	—	—	—	—	Granites belonging to this phase may underlie much of the batholith. Granites hydraulically fractured
Mb	First part of second phase of post-magmatic alteration and mineralization	275–270	? 2	450–380	Moderate	Mainly magmatic, some radiogenic	N–S or NW–SE	Greisenization: converted to quartz, mica and topaz by F-rich fluids, mica of gilbertite type. Tourmalinization: replaced by tourmaline	Repeatedly fractured and fractures annealed by fresh growths of quartz	Some re-crystallization, biotite loses iron which is taken up by tourmaline growth	Sn, W, Cu	Main phase of metalliferous mineralization
IIIc	Emplacement of felsitic elvan dykes	275–270	? 2	600–500	Moderate	Magmatic	N–S	—	—	—	Sn, W, Cu	Further input of magmatic heat

Iv	First phase of argillic alteration and N–SE or N–S quartz-hematite veins and faulting	270–260 ?	1–2	350–200	Moderate to high	Mainly radiogenic, possibly some magmatic or mantle heat I	E–W	Na feldspar: altered to smectite/illite assemblage, little kaolinite K feldspar: altered to illite, maybe some smectite	Free silica released by argillation, forms overgrowths on quartz and new iron-stained non-tourmaline bearing lodes (NW–SE and N–S)	Much iron liberated from biotite which is carried out of the granite to form iron lodes. Some mica hydrated to gilbertite	Fe/U/Pb/Zn	Note: Salinity, lack of kaolinite and change in stress direction. Low temperature metalliferous mineralization
Quiescent period?												
V	Second phase of argillic alteration. Main period of kaolinization ?Deep Mesozoic supergene alteration?	260 to present	0.2–1.5	50–150	Low	Radiogenic	Variable E–W or N–S, later becoming vertical	Na feldspar: altered readily to kaolinite K feldspar: altered less readily to kaolinite Smectite: altered readily to kaolinite	Free silica released by argillation, forms overgrowths on quartz and some minor quartz veins	Some iron liberated from biotite, not carried out of granite so colours matrix. In areas of intense kaolinization mica/illite altered to kaolinite	Fe/U (minor)	Note: Fresh water and main episode of kaolin formation. Isostatic uplift may have played a part

VI	Early Tertiary chemical weathering (also Mesozoic?)	25–60	0.0–0.2	20–50	Low	High surface temperature	Vertical	Altered kaolinite, is b-axis disordered in Eocene/Oligocene weathering	Some solution of silica from quartz grains	Some iron liberated from biotite, not carried out of the granite so colours matrix. In areas of intense kaolinization mica/illite altered to kaolinite	—	Tertiary weathering mantle is source of material for ball clays and associated sediments

Attention has been drawn to the K-rich nature of magmas associated with volcanic rocks in South-west England (Thorpe *et al.,* 1986; Leat *et al.,* 1987; Thorpe, 1987). These authors have deduced from isotopic compositions (for example, $[^{87}Sr/^{86}Sr]_{291}$ = 0.704–0.705; $[^{143}Nd/^{144}Nd]_{291}$ = 0.5123–0.5127) that they originated in a heterogeneous mantle which had previously undergone modification during a subduction event and that LIL-enriched mantle material might also have contributed to the granites. Leat *et al.* (1987) have proposed, largely on the basis of high concentrations of heat-producing elements, that potassic magmas originating in the mantle became contaminated by pelitic material to form a gabbroic, dioritic or minette magma body at the base of the crust, and that this then differentiated to form the main granite magma. The chief difficulties here seem to be the inappropriate isotope ratios of the granite (cf. (Table 2.1)) and the lack of unequivocal geophysical data; although not ruling out a sub-granite body of dioritic composition, the only calculated seismic velocity of 6.9 km s. would only just accommodate a body of such a composition (Holder and Bott, 1971). Of course, if the upper part of the batholith has been thrust away from its roots, as noted above, and suggested by the seismic reflection data of Brooks *et al.* (1984), such a body might exist somewhere to the south.

The second stage of magmatic activity, dated at about 270 Ma, was preceded by the development of a sheeted vein system in fissures opened in the carapace of the earlier granite. This was brought about by the build-up of pressure in a hydrothermal vapour with a low density and moderate salinity of 10–25 equiv. wt.% NaCl (Jackson *et al.,* 1982; Shepherd *et al.,* 1985). Its principal effect in rock terms, however, was the emplacement of a second magma which crystallized to give the albite–zinnwaldite–topaz granite exposed at St Austell, Tregonning and St Michael's Mount. Associated with the Li-mica granite intrusion at St Austell is a Li- and B-enriched variety of the earlier, main biotite granite generated as a metasomatic aureole, and also pockets of fluorite-bearing granite (Manning and Exley, 1984).

At the end of this stage, a series of roughly E–W, fine-grained, granite-porphyry dykes ('elvans') were intruded (Table 2.1). These are believed to have been derived from biotite granite magma, but in a number of cases show evidence of fluidization of solid material (Stone, 1968), indicating that some were not entirely magmatic. All the major plutons contain examples of fine-grained granite, but the origins and ages of these are not clear. Some are magmatic and intrusive, but others may be granitized sediment.

Although an early phase of mineralization, involving Fe–Mn and Fe–Cu, was related to pre-batholith sedimentary deposits (Jackson *et al.,* 1989), the main period of mineralization started at about 270 Ma and resulted from the build-up of hydrothermal fluids which were of high density and salinity (20–40 equiv. wt.% NaCl), and which, on entering the cover, deposited a series of roughly E–W-striking veins containing, for example, Sn, Cu, W and As minerals. The violence of such outbursts, accompanied by sudden decompression, sometimes produced spectacular breccias which are usually cemented by tourmaline. The field evidence as to the age of these breccias is somewhat equivocal, and in the absence of radiometric dating it is not always certain whether they were formed at the end of the first or the second magmatic phase.

Until the main stage of mineralization, argillic alteration was limited to greisening, some bordering mineral veins and some pervading the granites through a well-developed pore system, but temperatures, while fluctuating, probably never fell below about 380°C. Sheppard (1977) has shown that meteoric water was present during this process, emphazising the role played by groundwater from the country rocks in hydrothermal reactions.

Over the next 5–10 million years, as magmatic heat was lost, the temperature was maintained by radiogenic heat from K, Th and U, which are present in unusually high concentrations (Tammemagi and Smith, 1975), but nevertheless it fell slowly to an estimated 350–200°C. Fissures approximately at right angles to the earlier vein systems provided channels at this time for a further stage of mineralization which included Fe, Pb, Zn and U among other elements. The temperatures and salinities of the fluids at this time have been calculated at 200–150°C and 19–27 equiv. wt.% NaCI (Alderton, 1978; Shepherd *et al.,* 1985) and they gave rise to the first true clay minerals, namely smectite and illite, by alteration of the felsic minerals. The granite was thus prepared for subsequent kaolinization (Bristow, 1987; Bristow *et al.* in press).

The last-mentioned process has continued intermittently until the present day as a result of the circulation of water through the granites by convection cells driven by radiogenic heat (Durrance *et al.,* 1982; Durrance and Bristow, 1986; Bristow *et al.* in press). The reality of the circulation of meteoric water and its mixing with older water, made saline by reaction with plagioclase and biotite, has been confirmed by isotopic and selective cation analysis (Edmonds *et al.,* 1984). It has been established that the main radiogenic heat source resides in the accessory minerals apatite, zircon, monazite, xenotime and uraninite, of which the last is much the most significant (Ball and Basham, 1979; Jefferies, 1984, 1988). However, it has been suggested that additional heat derived from the mantle may also be required (Tammemagi and Smith, 1975), so invoking further speculation about the sub-batholith geology.

Sheppard (1977) has made it clear that the circulating water involved in the main process of kaolinization was meteoric, and it now seems likely that a period of Palaeogene subtropical weathering intensified the alteration already started (Bristow, 1988; Bristow *et al.* in press). The main stages of evolution are summarized in (Table 2.2).

Post-orogenic volcanic activity

After emplacement of the batholith and isostatic uplift towards the end of the Carboniferous, south-west England underwent rapid denudation during the Stephanian, Permian and Triassic that eventually unroofed the marginal facies of the granite plutons (Dangerfield and Hawkes, 1969). The eroded desert landscape was subsequently buried by aeolian sands and the pebbly sands and silts of flash floods and fluvial channels (Laming, 1966, 1968). Rapid lateral variations are common and in some cases reflect local contemporaneous faulting (Bristow and Scrivener, 1984). Locally derived debris within the basal conglomerates includes granitic material, aureole metasediments and acidic volcanics (Groves, 1931; Hutchins, 1963; Laming, 1966).

Within this 'red bed' continental environment during the late Carboniferous and probably early Permian period, various small-volume volcanic rocks were erupted. In terms of gross compositional variation, three main suites can be recognized – lamprophyres (limited occurrence throughout Cornubia), Exeter Volcanic 'Series' (restricted to the Crediton Trough and west of Exeter) and rhyolites (only seen at Kingsand, south of Plymouth).

Lamprophyre suite

Throughout south-west England is seen a set of NE–SW-trending minette-type lamprophyre dykes (Figure 2.8) that cut the Hercynian country-rock fabric and are considered to be post-orogenic in character (Hawkes, 1981; Darbyshire and Shepherd, 1985). These small intrusions are spatially associated with the granite, but apparently lie outside the surface outcrop (Exley *et al.,* 1982) and within the batholithic 'shadow zone' (Leat *et al.,* 1987). The post-orogenic nature of the lamprophyres, coming after major granitic emplacement, appears to be a global characteristic of some types of lamprophyric volcanism (Rock, 1984; Turpin *et al.,* 1988). Hawkes (1981) gave an average age of 291 Ma (whole-rock, K/Ar) for various lamprophyres that may be representative of this suite.

Exeter Volcanic 'Series'

Small-volume lava flows and dykes occur within the 'red bed' sequences to the north and west of Exeter (Figure 2.9), and on the basis of limited radiometric dates these are probably late-Stephanian in age. Although often considered to be Permian (Knill, 1982) with ages around 280 Ma (Miller and Mohr, 1964), the originally determined K/Ar dates have been recalculated to 291 Ma (Thorpe *et* al., 1986), and suggest some of the volcanics must be of very late Carboniferous age. Two main groups of volcanics are recognized – a basaltic suite and a potassic lava suite – that are collectively referred to as the Exeter Volcanic 'Series' (Knill, 1969, 1982; Cosgrove, 1972). The potassic suite is dominated by lamprophyres that are probably related to the minettes mentioned above and found throughout the region (Cosgrove and Hamilton, 1973; Exley *et al.,* 1982; Leat *et al.,* 1987).

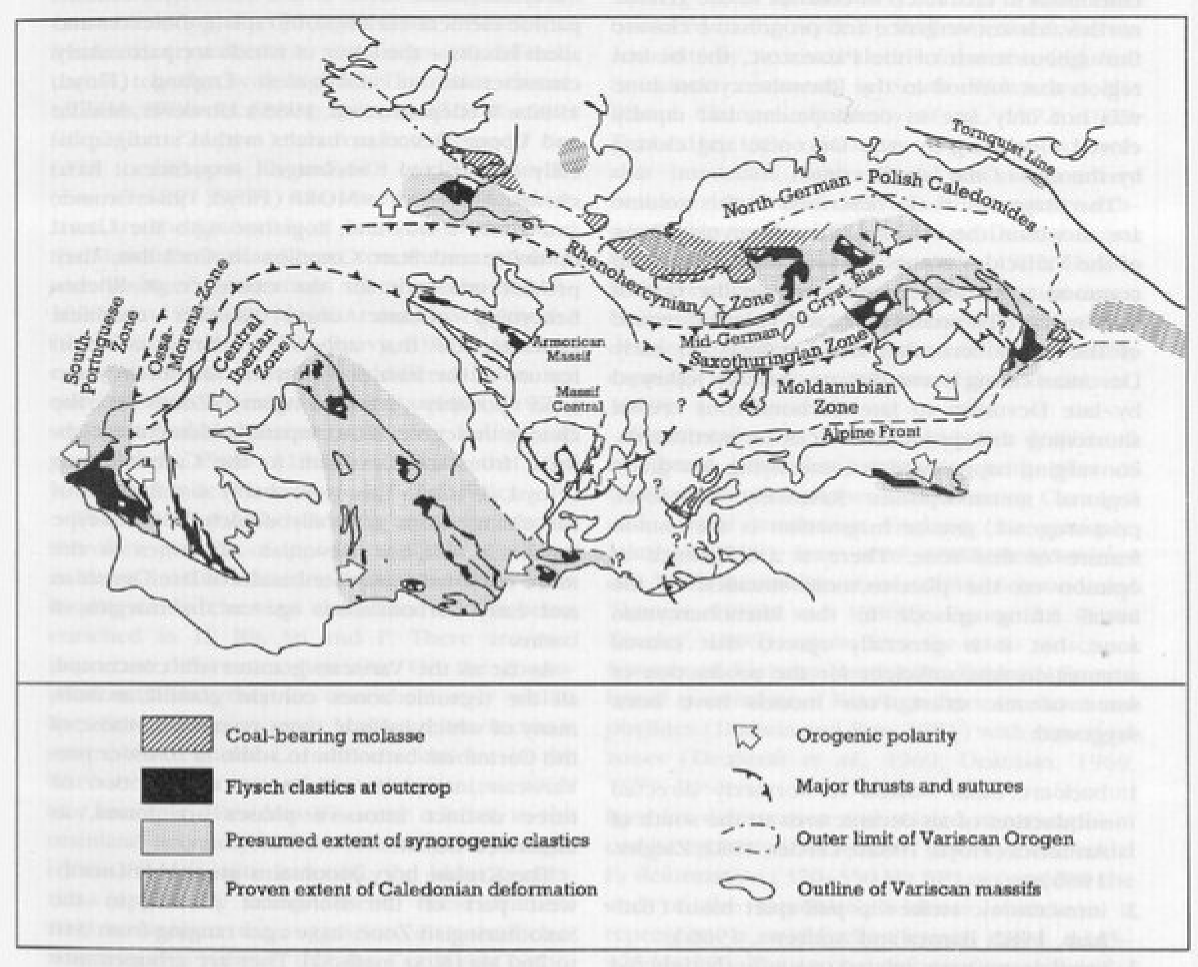
The Exeter Volcanic 'Series' was erupted into and over 'red bed' sequences and may show lava–soft-sediment mixing, as well as eroded lava surfaces infilled with sandstone (Knill, 1982). These volcanics are largely restricted to east–west trending grabens (for example, Crediton Trough, (Figure 2.9)) which developed during uplift after the Variscan Orogeny (Whittaker, 1975) or specifically via the reactivation of a major Variscan basement thrust by post-orogenic tension (Durrance, 1985b).

The significance of these volcanic rocks lies not only in their being representative of post-orogenic volcanism, but also in their relationship to the lamprophyres and the nature of the underlying mantle that supplied the mafic magmas. By far the most interesting feature concerns the potassic lavas; these exhibit a chemical fingerprint characteristic of subduction-related magmas which suggests they were derived from lithosphere subducted during the Variscan (Thorpe *et al.,* 1986; Leat *et al.,* 1987; Thorpe, 1987). Although this does not necessarily imply the presence of an active subduction zone during the Permian (for which there is no direct geological evidence), Cosgrove (1972) originally suggested that the whole 'shoshonitic suite', as he referred to the Exeter Volcanic 'Series', was generated at a plate margin.

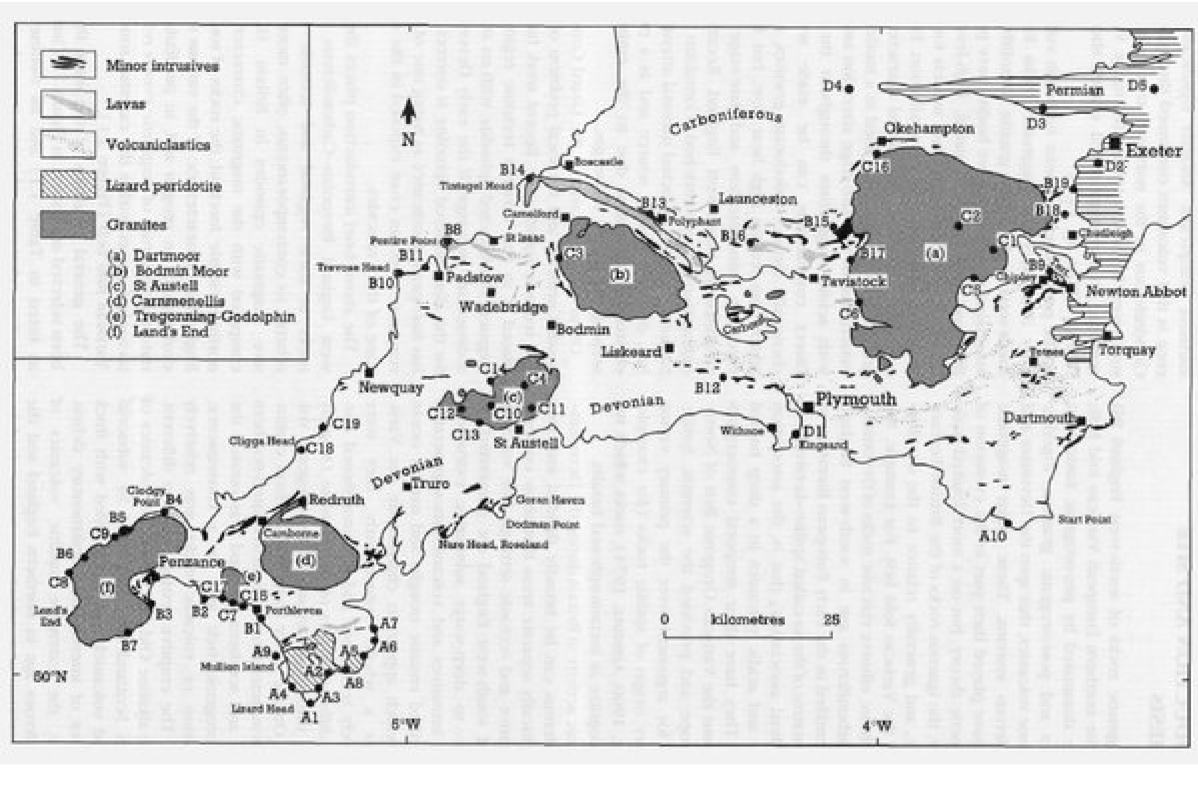
Rhyolitic lava group

Remnants of flow-banded porphyritic rhyolitic lavas, apparently lying on deformed Devonian argillites, occur at Kingsand (near Plymouth, (Figure 1.1)) and also as numerous pebbles within the 'red bed' conglomerates (Laming, 1966). Both occurrences form a comagmatic suite (Cosgrove and Elliott, 1976) and are considered to be representative of widespread subaerial acid volcanism fed by late-stage dykes (Hawkes, 1974). The rhyolites are not apparently related to the granites but have petrographic and chemical affinities with the late, granite-porphyry dykes (Exley *et al.,* 1983) that were emplaced as fluidized systems during the 280–270 Ma interval (Exley and Stone, 1982; Darbyshire and Shepherd, 1985).

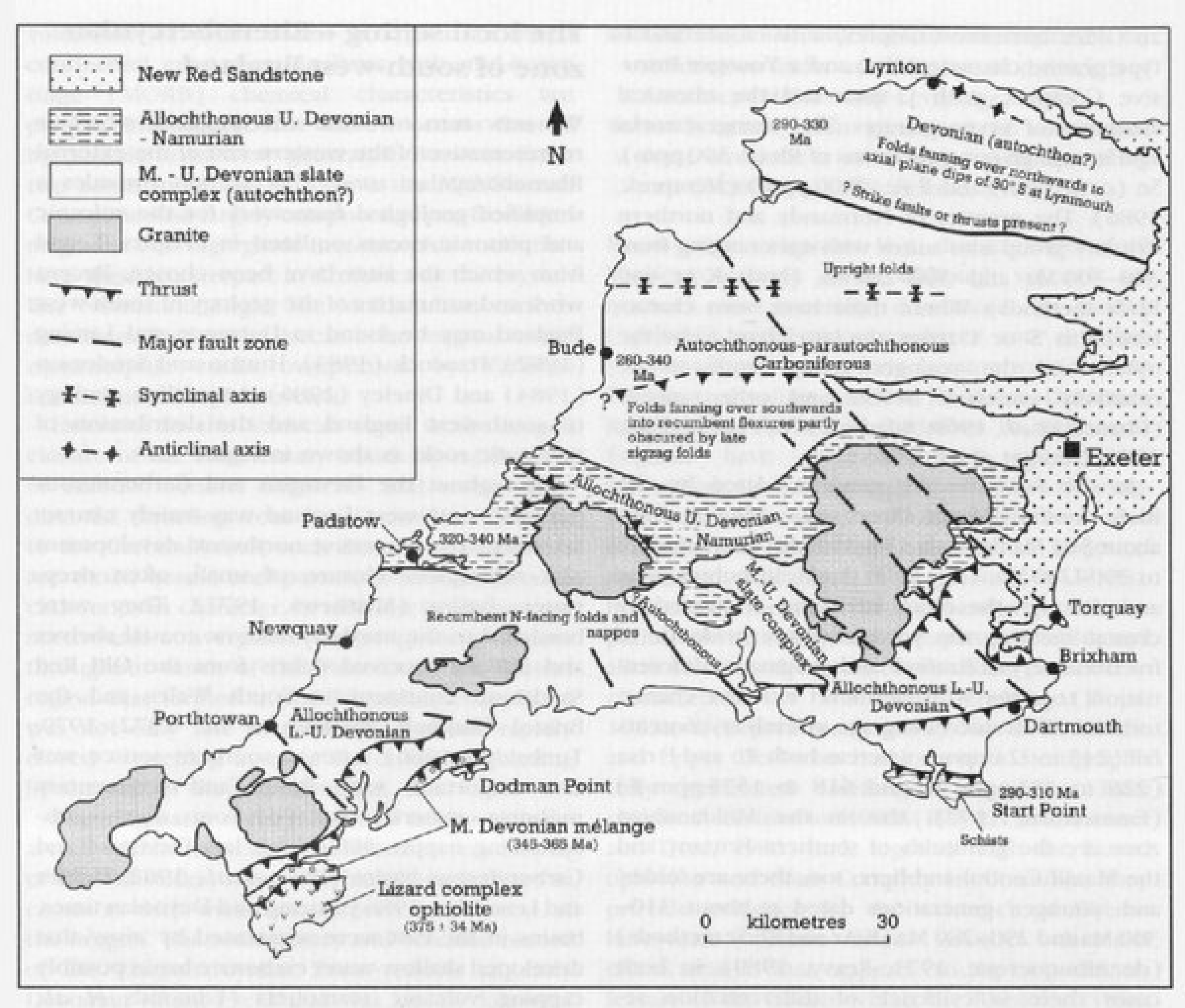
References



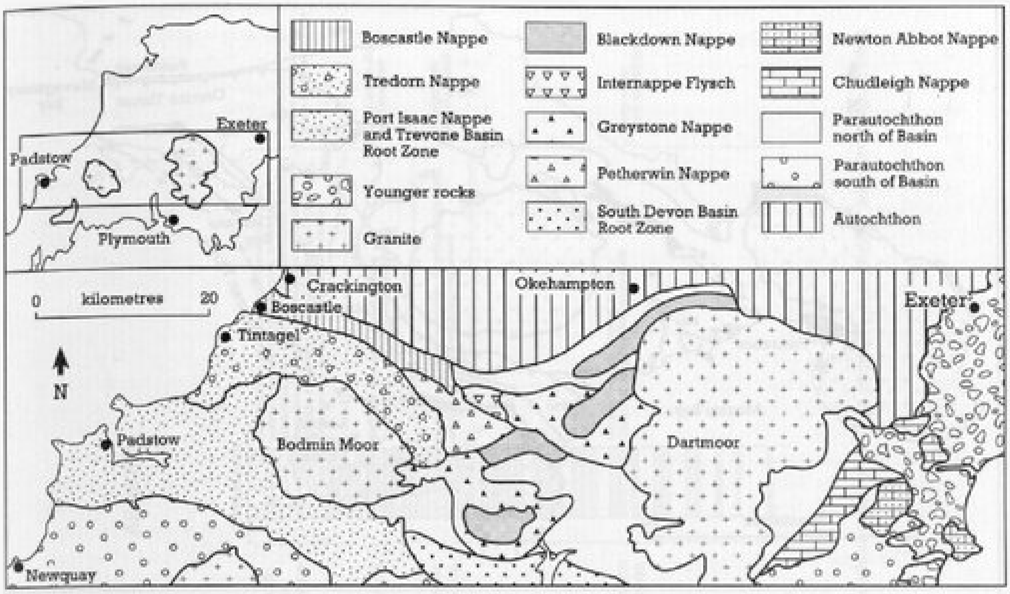
(Figure 2.1) Distribution of tectonic zones in the Variscan Orogen of Europe (modified from Franke, 1989).



(Figure 1.1) Simplified geological map of south-west England showing the distribution of magmatic rocks and the approximate location of sites described in the text (modified from Floyd, 1982b). Sites are numbered and grouped as in (Table 1.1).

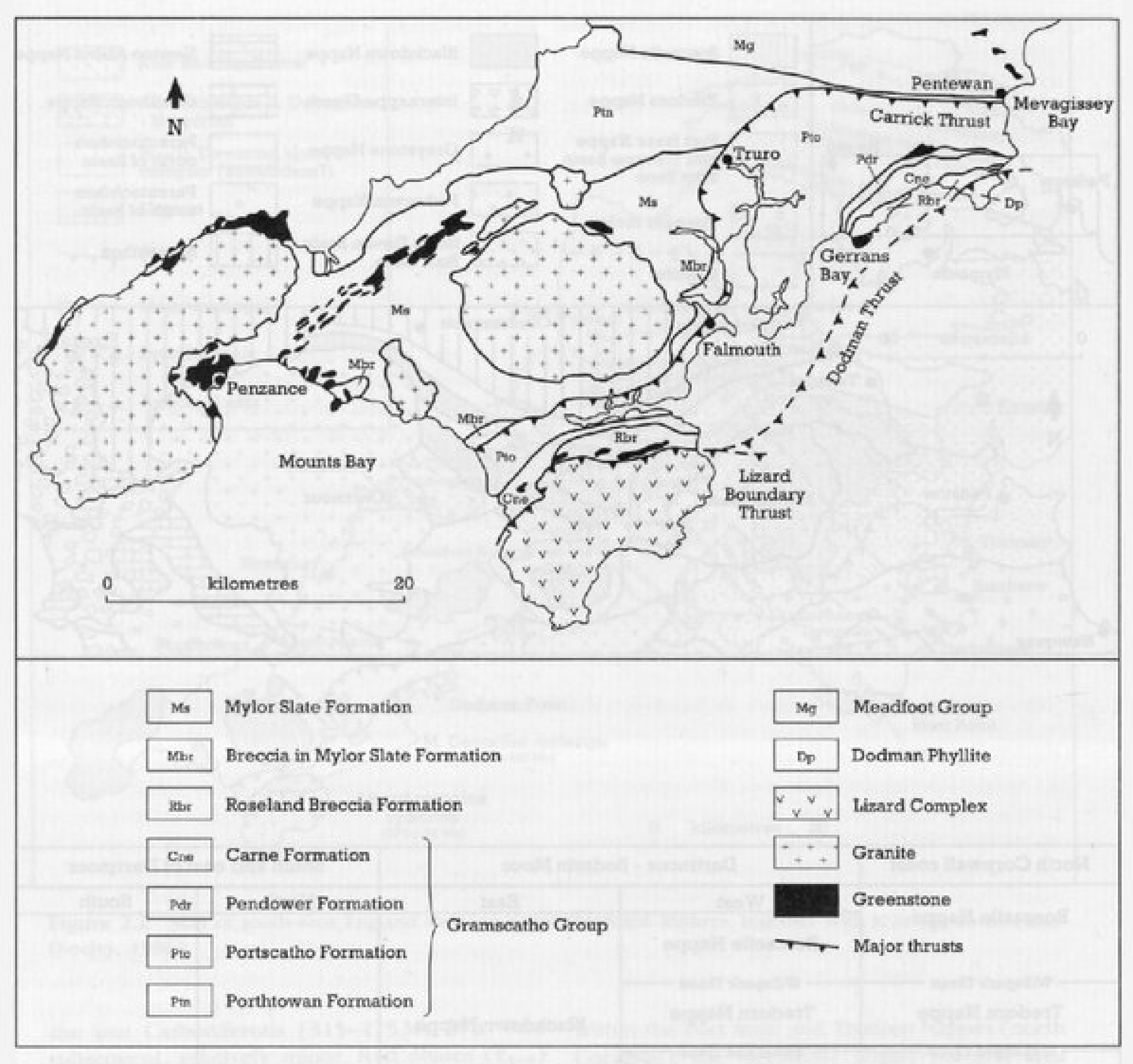


(Figure 2.2) Map of south-west England showing major structural features, together with K/Ar age zones (after Dineley, 1986).

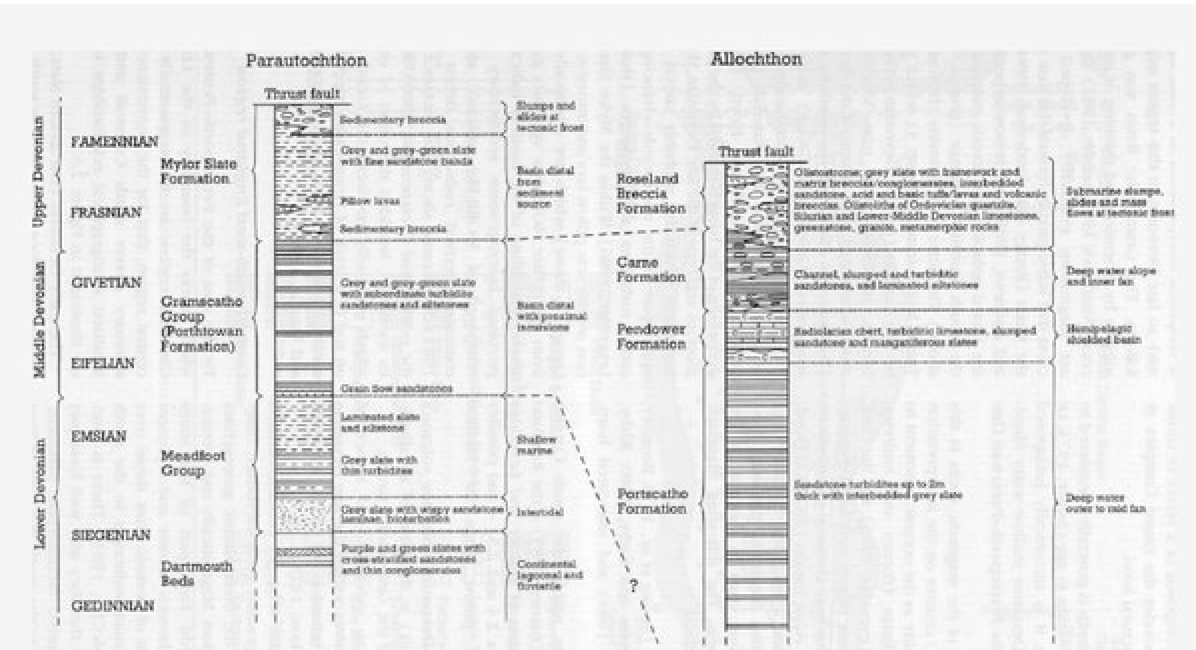


North Cornwall coast	Dartmoor - Bodmin Moor		South and east of Dartmoor	
	West	East	North	South
Boscastle Nappe	Boscastle Nappe	Blackdown Nappe		
Willapark Thrust	Willapark Thrust			
Tredorn Nappe	Tredorn Nappe			
Trekelland Thrust	Trekelland Thrust			
Port Isaac Nappe	Petherwin Nappe Greystone Thrust Greystone Nappe	Blackdown Thrust Greystone Nappe	Chudleigh Nappe	Newton Abbot Nappe
			Bickington Thrust	Forder Green Thrust
Thrust	Main Thrust	Main Thrust		
Parautochthon ? Kate Brook Unit	Parautochthon Kate Brook Unit	Parautochthon Kate Brook Unit	Parautochthon Teign Valley and Kate Brook Units	Parautochthon Denbury Unit

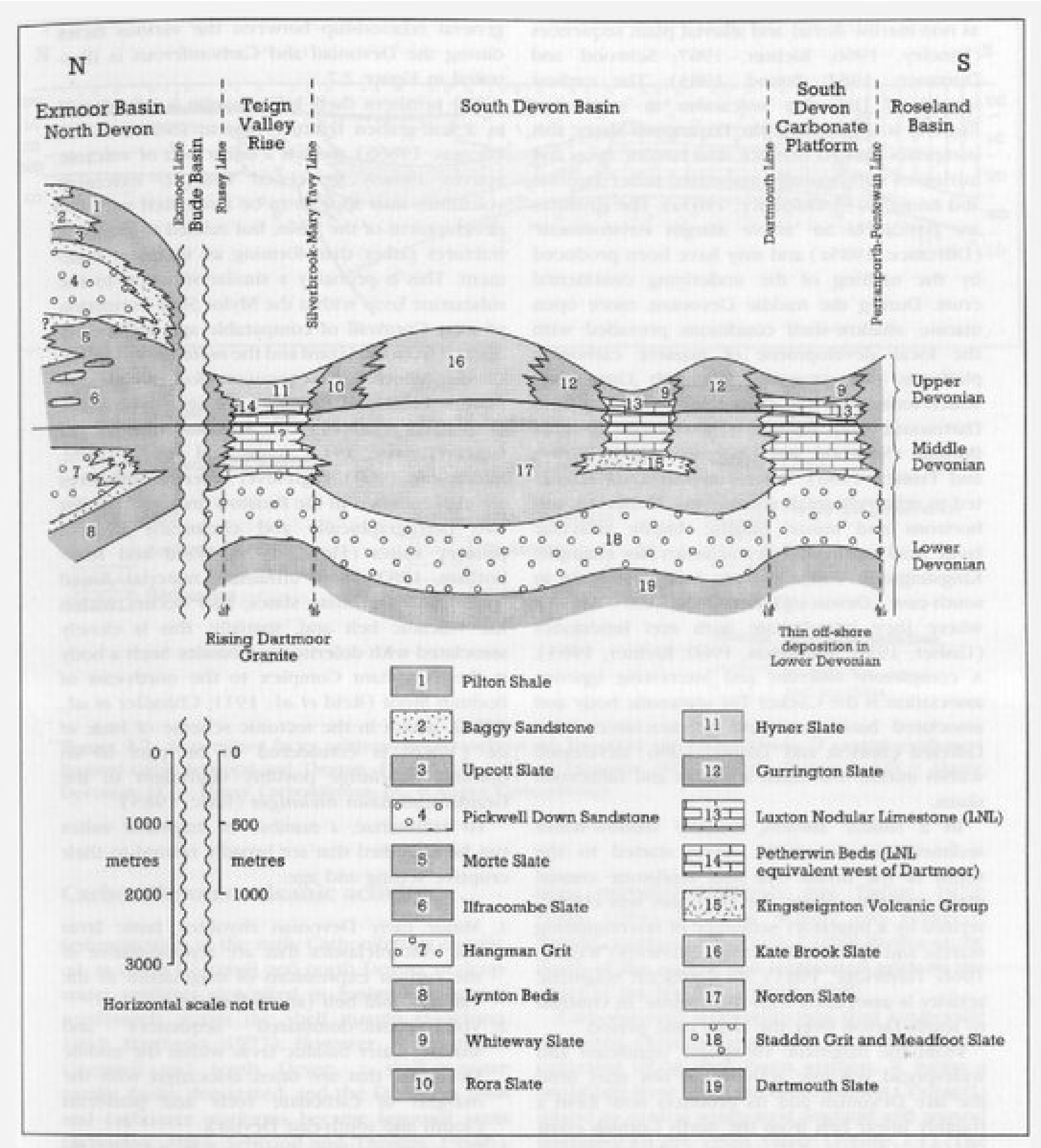
(Figure 2.3) Distribution and correlation of structural units in central south-west England (after Selwood and Thomas, 1986a).



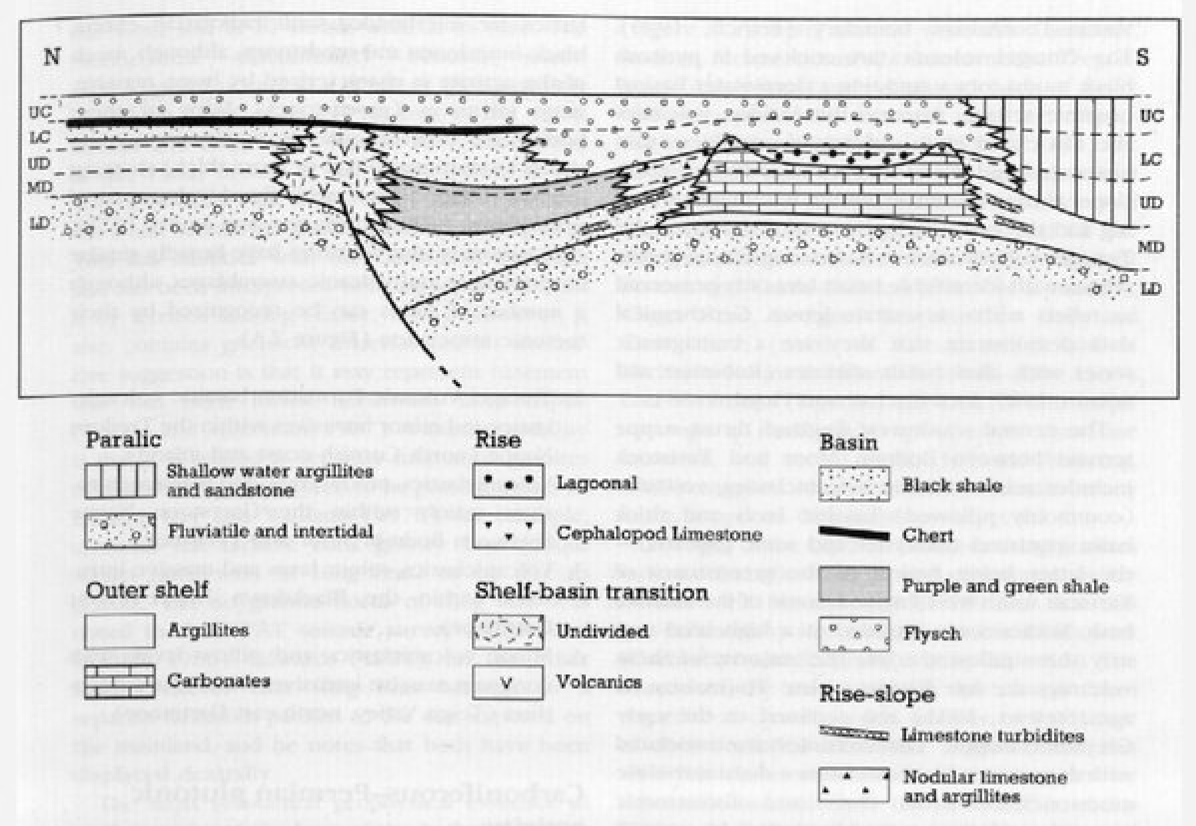
(Figure 2.4) Geological map of south Cornwall showing various allochthonous units, including the Lizard Complex, resting on the northern parautochthon of the Porthtowan Formation and Mylor Slate Formation (after Holder and Leveridge, 1986).



(Figure 2.5) Devonian lithostratigraphical sequences in the parautochthon and allochthon units of south Cornwall (after Holder and Leveridge, 1986).



(Figure 2.6) Correlation of observed Devonian lithostratigraphy across Devon (after Durrance and Laming, 1982).



(Figure 2.7) Generalized facies relationships throughout the Devonian and Carboniferous of central south-west England (after Selwood and Thomas, 1986b). LD = Lower Devonian; MD = Middle Devonian; UD = Upper Devonian; LC = Lower Carboniferous; UC = Upper Carboniferous.

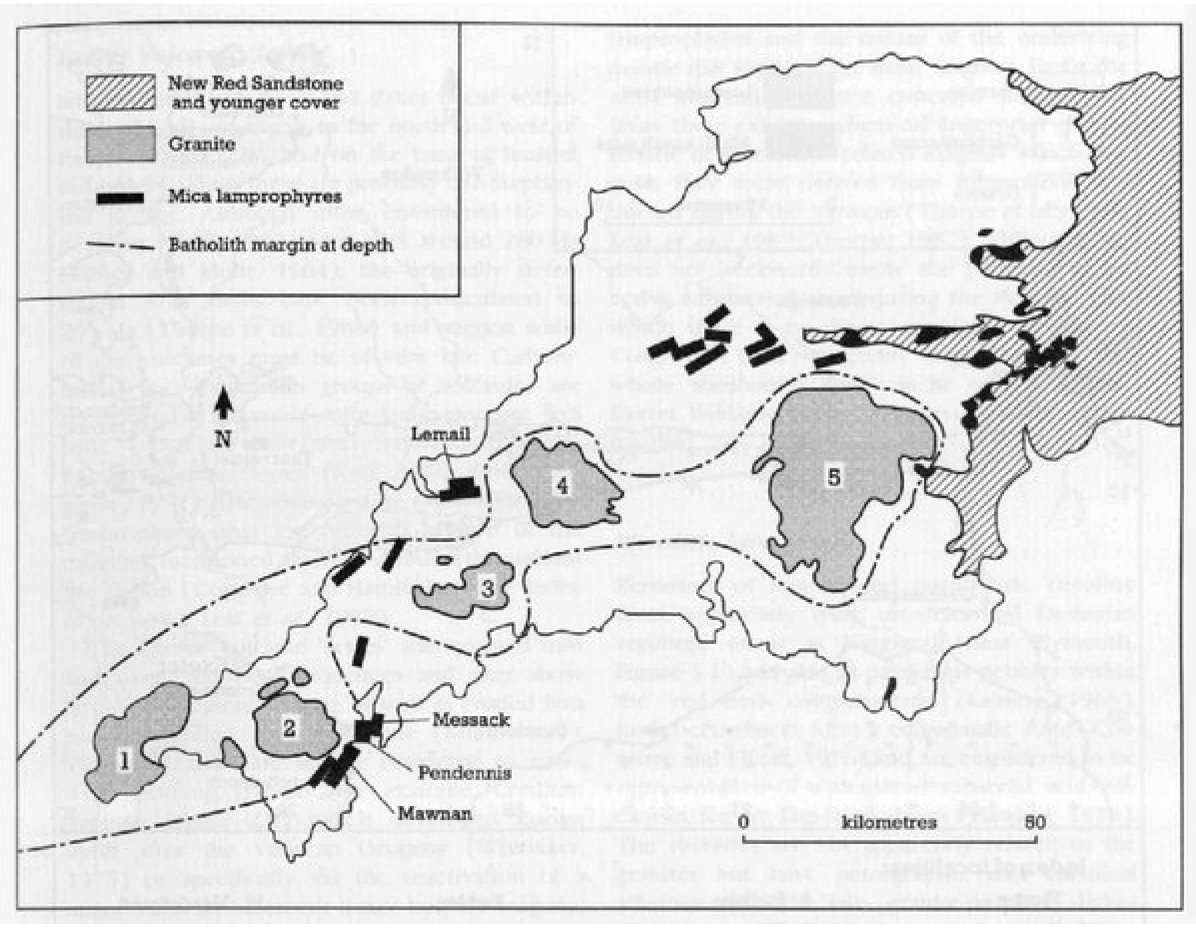
Intrusive phase	Outcrop and granite type	Rb-Sr age (Ma)	Initial $^{87}Sr/^{86}Sr$ ratio	Comments
Major	Dartmoor (B)	280 ± 1	0.7101 ± 0.0004	-
	Bodmin Moor (B)	287 ± 2	0.7140 ± 0.0002	Mineral age
	St Austell (B)	285 ± 4	0.7095 ± 0.0009	-
	Carnmenellis (B)	290 ± 2	0.7130 ± 0.0020	Mineral age
	Tregonning (E)	280 ± 4	0.71498 ± 0.00381	Highly evolved, lithium-rich
	Land's End (B)	268 ± 2	0.7133 ± 0.0006	Mineralization re-set age
Minor	Hemerdon Ball	304 ± 23	0.70719 ± 0.01025	Heavily mineralized
	Kit Hill	290 ± 7	0.70936 ± 0.00228	-
	Hingston Down	282 ± 8	0.71050 ± 0.00119	-
	Castle-an-Dinas	270 ± 2	0.71358 ± 0.00122	Later intrusion re-set age
	Carn Marth	298 ± 6	0.70693 ± 0.00207	-
Dykes	Meldon 'Aplite'	279 ± 2	0.7098 ± 0.0017	-
	Brannel Elvan	270 ± 9	0.7149 ± 0.0031	Re-analysed
	Wherry Elvan	282 ± 6	0.7120 ± 0.0025	Re-analysed
Mineral veins	South Crofty	269 ± 4	-	-
	Geevor	270 ± 15	0.7122 ± 0.0012	-

(Table 2.1) Ages and initial Sr isotopic ratios of granitic rocks from the Cornubian batholith (data from Darbyshire and Shepherd, 1985, 1987)

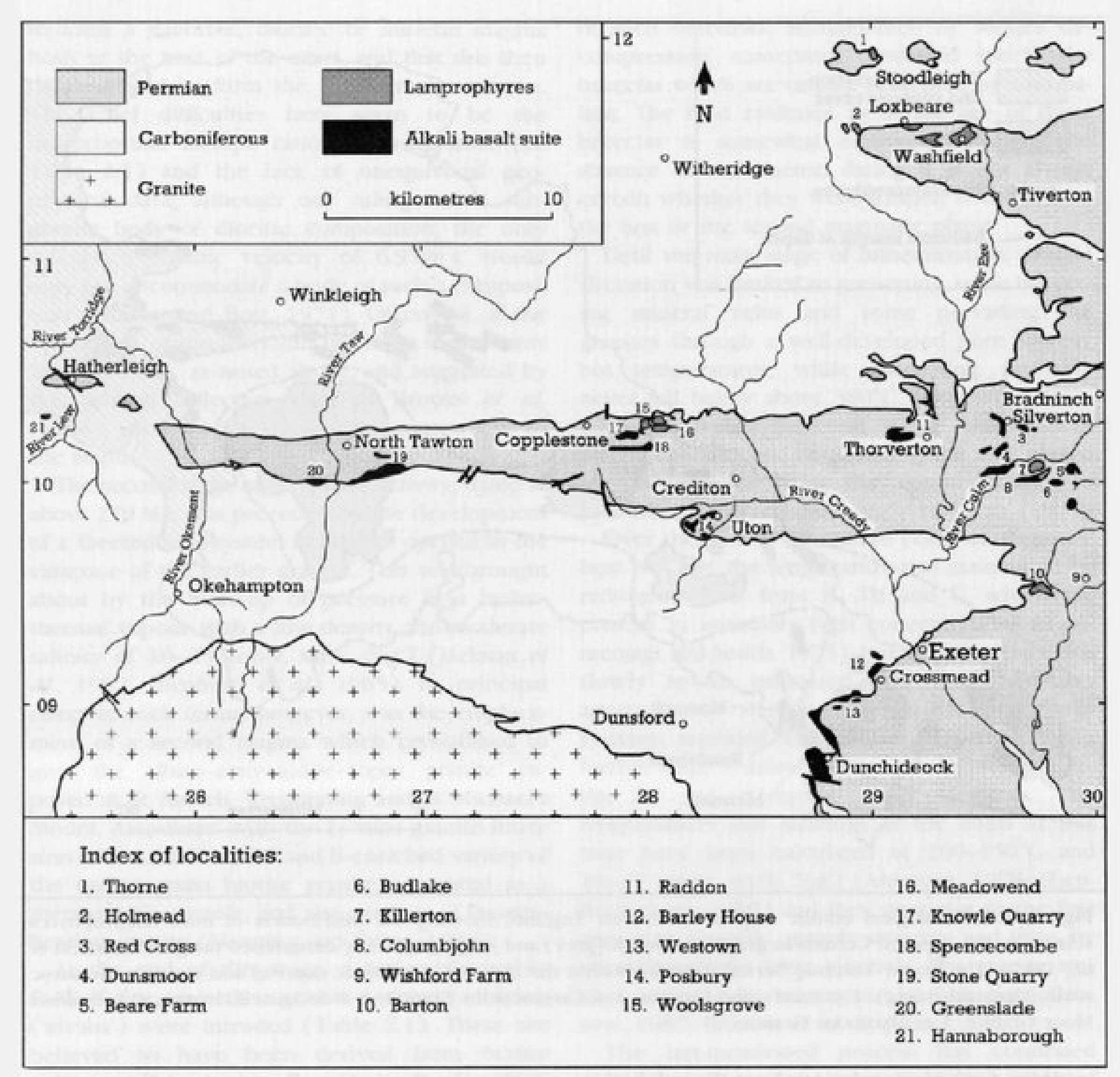
Stage	Process	Age (millions of years) *	Depth (km)	Temperature (°C)	Salinity of fluids	Source of heat	Direction of least stress	Main changes in mineralogy			Associated metalliferous mineralization	Comments
								Feldspar	Quartz	Mica		
I	Emplacement of biotite granite, forming main batholith	290-285	? 3	500-600	-	Magmatic	Variscan (E-W)	-	-	-	-	Biotite granite which now forms eastern part of the St Austell granite
II	First phase of post-magmatic alteration and mineralization	285-275	2-3	500-200	Moderate	Magmatic	Initially E - W, then N - S	Limited greisenization alongside veins	-	-	Sn, W	Early greisenization and mineralization e.g. Castle-an-Dinas (W)
IIIa	Emplacement of evolved lithium-rich granites and biotite granites in western part of St Austell granite	275-270	2-3	500-600	-	Magmatic	N - S	-	-	-	-	Granites belonging to this phase may underlie much of the batholith. Granites hydraulically fractured
IIIb	First part of second phase of post-magmatic alteration and mineralization	275-270	? 2	450-380	Moderate	Mainly magmatic, some radiogenic	N - S or NW - SE	Greisenization: converted to quartz, mica and topaz by F-rich fluids, mica of gilbertite type. Tourmalinization: replaced by tourmaline	Repeatedly fractured and fractures annealed by fresh growths of quartz	Some re-crystallization, biotite loses iron which is taken up by tourmaline growth	Sn, W, Cu	Main phase of metalliferous mineralization
IIIc	Emplacement of felsitic elvan dykes	275-270	? 2	400-500	Moderate	Magmatic	N - S	-	-	-	Sn, W, Cu	Further input of magmatic heat
IV	First phase of argillic alteration and NW-SE or N-S quartz-hematite veins and faulting	270-260	? 1-2	350-200	Moderate to high	Mainly radiogenic, possibly some magmatic or mantle heat	E - W	Na feldspar: altered to smectite/illite assemblage, little kaolinite K feldspar: altered to illite, maybe some smectite	Free silica released by argillation, forms overgrowths on quartz and now iron-stained non-tourmaline bearing lodes (NW - SE and N - S)	Much iron liberated from biotite which is carried out of the granite to form iron lodes. Some mica hydrated to gilbertite	Fe/U/Pb/Zn	Note: Salinity, lack of kaolinite and change in stress direction. Low temperature metalliferous mineralization
Quiescent period?												
V	Second phase of argillic alteration. Main period of kaolinization ?Deep Mesozoic supergene alteration?	260 to present	0.2-1.5	50-150	Low	Radiogenic	Variable E - W or N - S, later becoming vertical	Na feldspar: altered readily to kaolinite K feldspar: altered less readily to kaolinite Smectite: altered readily to kaolinite	Free silica released by argillation, forms overgrowths on quartz and some minor quartz veins	Some iron liberated from biotite, not carried out of granite so colours matrix. In areas of intense kaolinization mica/illite altered to kaolinite	Fe/U (minor)	Note: Fresh water and main episode of kaolin formation. Isostatic uplift may have played a part
VI	Early Tertiary chemical weathering (also Mesozoic?)	25-60	0.0-0.2	20-50	Low	High surface temperature	Vertical	Altered kaolinite, is b-axis disordered in Eocene/Oligocene weathering	Some solution of silica from quartz grains	Some iron liberated from biotite, not carried out of the granite so colours matrix. In areas of intense kaolinization mica/illite altered to kaolinite	-	Tertiary weathering mantle is source of material for ball clays and associated sediments

* Radiometric dates from Bray (1980), and Darbyshire and Shepherd (1985, 1987)

(Table 2.2) Main evolution and alteration stages of the St Austell Granite (after Bristow et al., in press)



(Figure 2.8) Geological outline map of south-west England, showing the distribution of mica lamprophyres relative to the exposed Cornubian granite batholith (grey) and its geophysically determined margin (after Leat et al., 1987). The Exeter Volcanic 'Series' is shown within the New Red Sandstone outcrop and younger Mesozoic rocks (diagonal ruling). 1 = Land's End Granite; 2 = Carnmenellis Granite; 3 = St Austell Granite; 4 = Bodmin Moor Granite; 5 = Dartmoor Granite.



(Figure 2.9) Distribution of the two main magmatic groups within the Exeter Volcanic 'Series', mid-Devon (after Edmonds et al., 1969).