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ALTERATION AND VEIN MINERALISATION WITHIN THE LIZARD COMPLEX, SOUTH CORNWALL: CONSTRAINTS ON THE TIMING OF SERPENTINISATION.

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Two distinct episodes of serpentinisation have been identified within the peridotites of the Lizard complex. The first episode (primary serpentinisation) is represented by the complex and pervasive hydration of the Lizard peridotite. The second (later) episode is characterised by a pale to dark green, pseudo-fibrous mixture of lizardite and chrysotile that is restricted to fractures (vein serpentine). Mineralised north-north-west and east-north-east trending fault zones contain fragments of vein serpentine generated during this second episode. Faults with similar orientations and mineralogy within the gabbro unit contain adularia which have been previously dated by Ar^{40}-Ar^{39} and K-Ar methods at 210-220 Ma (Triassic). Stable isotope ratios indicate that the mineralisation within the gabbro and peridotite is genetically similar. The primary and secondary serpentinisation episodes are therefore interpreted as pre-Triassic in age. A latest Carboniferous to early Permian age is proposed for the formation of vein serpentine and a late Devonian to Carboniferous age is proposed for the primary serpentinisation episode. This is envisaged to have taken place post-obduction but a pre-obduction initiation of serpentinisation cannot be discounted.

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INTRODUCTION

The Lizard complex of south Cornwall (Figure 1) comprises variably metamorphosed peridotites, gabbros and basalts which are associated with acid/intermediate/basic gneisses and metasediments (e.g. Flett, 1946). The peridotites, gabbros and basalts represent oceanic lithosphere that was generated during Devonian rifting and subsequently obducted during Variscan convergence (e.g. Bromley, 1979; Kirby, 1979; Floyd *et al.,* 1993). These lithologies have undergone a complex alteration and mineralisation history. It has long been recognised that the peridotites display extensive pervasive serpentinisation (e.g. Bonney, 1877), and the serpentine group mineral lizardite was first described from the Lizard (Midgley, 1951), but there have no detailed studies. In addition, fracture-hosted mineralisation occurs in all lithologies, but has only been recorded in detail from the gabbros (e.g. Seager, 1971). The purpose of this contribution is to present new field observations regarding: (i) the nature of pervasive serpentinisation, (ii) the mineralogy and paragenesis of fracture-hosted mineralisation, and (iii) correlation of fracture hosted mineralisation between gabbros and peridotites. These data are used, in combination with previously published geochronological data, to provide preliminary constraints upon the timing of pervasive serpentinisation.

LIZARD PERIDOTITES

The Lizard complex peridotites are exposed over an area exceeding 30 km^2 (Figure 1) and comprise variably tectonised and hydrated harzburgites, lherzolites and dunites (e.g. Flett, 1946; Green, 1964a). Strictly speaking, all of these lithologies are serpentinites or partially serpentinised peridotites, but due to historical precedence (e.g. Green, 1964a, b) they are classified on the basis of their primary mineralogy. Three types of peridotite were defined by Green (1964a) on the basis of mineral assemblage: the primary, the recrystallised anhydrous, and the recrystallised hydrous assemblages. The primary assemblage comprises olivine, Al-rich enstatite, chrome-rich clinopyroxene and an olive green, aluminous spinel (Green, 1964a). This assemblage exhibits a coarse anhedral texture and is characterised by large orthopyroxene porphyroclasts. Relict cumulate and crescumulate textures have been identified within the assemblage by Rothstein (1977, 1981, 1994). A similar mineral assemblage is present in the recrystallised anhydrous peridotite, however plagioclase and low-Al pyroxenes are present and there is some replacement of clinopyroxene by hornblende (Green, 1964a). A granoblastic texture is evident, with prominent porphyroclasts of orthopyroxene and spinel (Green, 1964a). The recrystallised hydrous assemblage comprises olivine, pargasite and chrome spinel (Green, 1964a); recrystallisation has resulted in marked grain size reduction (crystals typically <1mm) and a pronounced foliation is defined by the pargasite (Green, 1964a). The differences between these assemblages are gradational (e.g. Rothstein, 1981; Floyd *et al.,* 1993) and may be interpreted in terms of variable degrees of recrystallisation and hydration in a broadly retrograde P-T environment associated with mantle diapir emplacement (Green, 1964a; Styles and Kirby, 1980), or the proximity to a major thrust fault (the Goonhilly Downs Thrust; Power *et al.,* 1996).

SERPENTINISATION

Processes

The silicate minerals of ultramafic rocks are very susceptible to hydration and are invariably altered to minerals of the serpentine group ($Mg_3Si_2O_5(OH)_4$), typically lizardite and chrysotile or, more rarely, antigorite. Lizardite and chrysotile develop at temperatures between 20-250°C (e.g. Barnes and O'Neil, 1969; O'Hanley, 1996). Antigorite develops at higher temperatures, typically in the range 250-400°C (e.g. O'Hanley, 1996) and is usually associated with regionally metamorphosed ultramafic bodies. Fe is not readily accepted within the lattice of serpentine group minerals, and magnetite is always formed as a by-product of serpentinisation. The abundance of magnetic

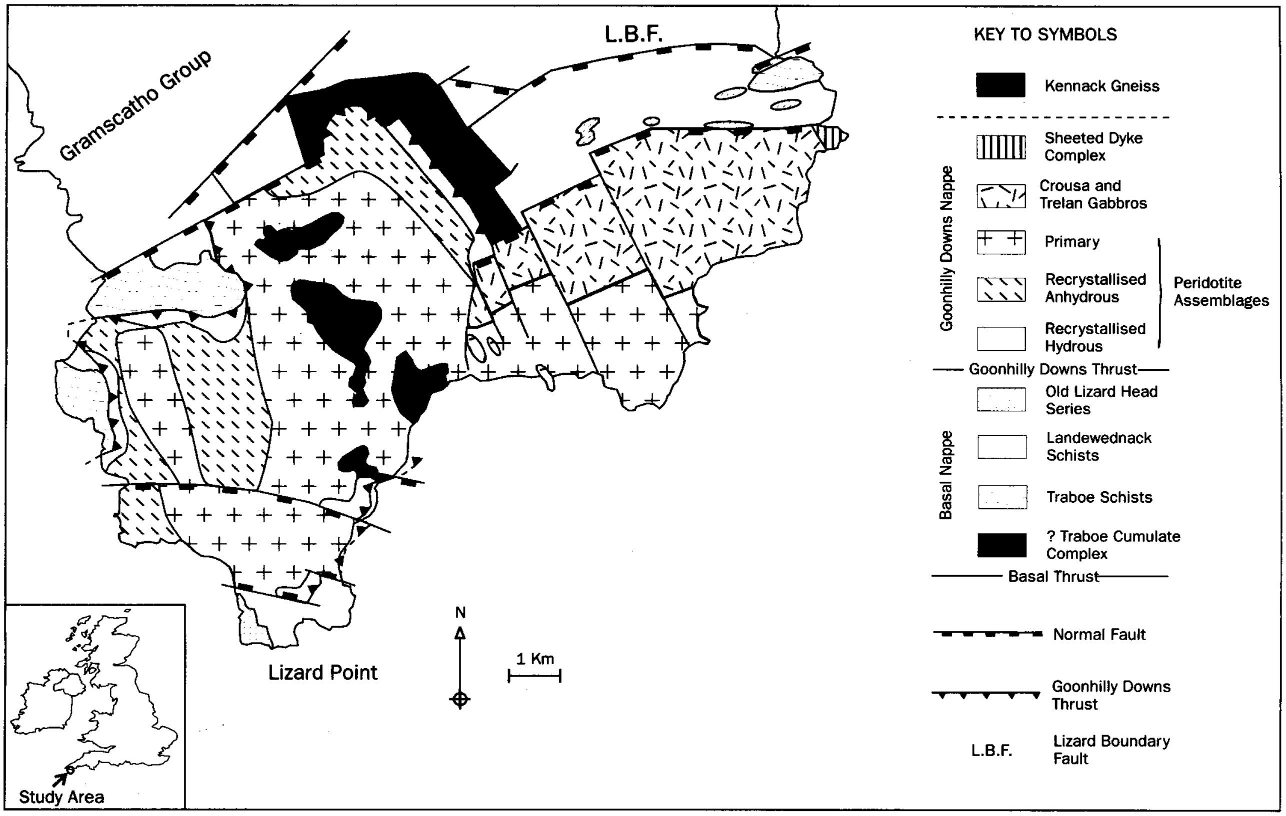

Figure 1: A simplified geological map of the Lizard complex showing the distribution of peridotite lithologies in relation to the Goonhilly Downs Thrust. Modified after Green, 1964; Leake et al., 1992; Power et al., 1996.

increases with increasing temperature, since at low temperatures, typically within the lizardite-chrysotile stability field, some Fe partitions into brucite (Moody, 1976). Serpentine minerals may also occur as precipitates in fractures; this material (vein serpentine) is typically translucent to green, pseudo-fibrous and free from magnetite inclusions (e.g. O'Hanley, 1996).

Olivine is significantly denser than serpentine minerals, hence a volume increase in the order of 50% is anticipated during constant mass serpentinisation (O'Hanley, 1996). Structures have been identified by O'Hanley (1992) where "kernels" of partly serpentinised rock are enclosed by rims of totally serpentinised rock containing serpentine-filled expansion cracks. These kernel structures are interpreted as a product of the mechanism by which expansion is accommodated during serpentinisation (O'Hanley, 1992). If serpentinisation does not conserve mass (i.e. in an open system) then it must be accompanied by removal of material in solution (O'Hanley, 1992). Localised Ca metasomatism commonly accompanies the serpentinisation process giving rise to rodingites; a typical rodingite mineral assemblage comprises idocrase, grossular, prehnite, diopside and chlorite (Coleman, 1977). However, whilst Ca metasomatism is widely reported, Si and, to a lesser degree, Mg appear to be retained indicating that loss of mass during serpentinisation is minimal (O'Hanley, 1992).

Previous studies on the Lizard

The serpentinisation of the Lizard peridotite has been widely reported (e.g. Bonney, 1877; Flett, 1946) but there has been no detailed investigation of the nature or timing of this process. Hydration of the ferromagnesian phases has given rise to a serpentine assemblage of lizardite and chrysotile (Midgley, 1951; Green, 1964b); antigorite is uncommon and restricted to zones that are in close proximity to the Goonhilly Downs Thrust (e.g. Green, 1964a, b; Power *et al.,* 1996).

Serpentinisation is variable across the complex (although rarely complete) with the least altered samples being approximately 30% serpentinised (Floyd *et al.,* 1993). However, since olivine is more susceptible to hydration than the other ferromagnesian phases (e.g. Wicks, 1969), the dunites are commonly totally serpentinised (Floyd *et al.,* 1993).

Rodingites have been described from Enys Head [SW 728 149] by Hall (1979) and Polbream Point [SW 730 158] by Green (1964a). These are pale green, have a saccharoidal texture, and possess a typical assemblage of grossular, chlorite, and diopside (Hall, 1979). Features also thought to be related to rodingite alteration include the development of prehnite and amphibole in the basaltic dykes which cut the peridotite (Hall, 1979); and the formation of massive talc, from granite, at Gew Graze [SW 676 144] (Hall, 1979). Although rodingites have been described from the complex, they are volumetrically insignificant, leading Green (1964a) to believe that mass was essentially conserved during serpentinisation of the Lizard peridotites. Flett (1946) also noted that, in troctolites, numerous fractures radiate out from weathered olivine grains and displace plagioclase crystals; these fractures were attributed to expansion of the olivine during serpentinisation.

The serpentinisation process was believed by Flett (1946) to be a protracted weathering process which was still progressing. Floyd *et al.* (1993) also presumed the process to be continuing at present, but they proposed that it initiated during obduction at high temperature (400-500°C) with the formation of coarse lizardite. In the discussion in Hall (1979) it was noted that the rodingite alteration affects Kennack Gneiss (Figure 1). The Kennack Gneiss is a group of variably deformed acid to basic gneisses and associated granitic dykes (e.g. Pearce, 1989) which are interpreted as a composite intrusion of basic and acidic magmas injected along the Goonhilly Downs Thrust during the early stages of obduction (Sandeman, 1988). Accordingly, rodingite alteration is unlikely to have initiated prior to the onset of obduction (Hall, 1979).

Field Relationships

Two episodes of serpentinisation are evident from field relationships. The first is a complex episode responsible for the variable pervasive hydration of the peridotites to a serpentine assemblage of lizardite, chrysotile and magnetite. The second, volumetrically minor episode, is represented by laterally persistent veins comprising lizardite and chrysotile (vein serpentine).

Kernel structures, developed during the pervasive serpentinisation event, are present throughout the Lizard peridotite and are particularly well developed at southern Kennack Sands [SW 734 164]. They are characterised by veins of white serpentine (usually one main vein tangential to the centre of the kernel with short veins orientated perpendicular to it) enclosing a core of less serpentinised peridotite. Orthopyroxene crystals are lighter in colour towards the rims of the kernels; the rims often follow preexisting anisotropies (e.g. rodingitised gabbro dykes). Veins of magnetite are frequently present within the rims of the kernel structures and are particularly prominent in the recrystallised anhydrous assemblage and dunites.

Vein serpentine is abundant across the complex and can be distinguished from the serpentine associated with kernel structures by the pseudo-fibrous habit and lateral persistence of the infill. In thin section the material is free of magnetite inclusions and has a laminated habit characteristic of crack-seal processes (Figure 2). Where it is in contact with serpentinised peridotite there is no evidence of recrystallisation of the vein serpentine; its lamination is preserved. Dendritic masses of native copper, containing small amounts of native silver, are commonly found in association with vein serpentine (e.g. Trevassack quarry [SW 712 222]).

Although not widely documented, minor rodingites are abundant within the Lizard complex peridotites. They appear to be derived from localised Ca metasomatism of thin gabbroic dykes and typically comprise grossular, diopside, chlorite and purple idocrase. The rodingites are invariably cut by small serpentine veins related to kernel structures (and occasionally vein serpentine), usually perpendicular to the rodingite; no preferred orientation of rodingites has been observed. In addition to the widely documented talc deposit at Gew Graze (which is no longer exposed; e.g. Flett, 1946), massive talc occurs at a number of other localities (e.g. Kynance Cove [SW *685 133]*, north Pentreath [SW 692 129]). It occurs as pods and screens in close proximity to acid fractions of Kennack Gneiss and contains numerous Fe-rich relicts which bear a remarkable resemblance (both in size and density of distribution) to orthopyroxene crystals.

FRACTURE-HOSTED MINERALISATION

Structural framework

The peridotites, gabbros and basalts of the Lizard complex have been successively deformed during a series of tectonic episodes including: (i) ridge-axis or near-ridge extension (Gibbons and Thompson, 1991; Roberts *et al*., 1993), (ii) northwesterly imbrication and obduction during late Devonian convergence (e.g. Rattey and Sanderson, 1984; Jones, 1994) and (iii) post-obduction extensional and strike-slip faulting (Alexander and Shail, 1996; Power et al., 1996). The mineralisation described below is predominantly hosted by east-north-east to east-southeast and north-north-west striking high angle regional fracture sets. At outcrop scale, both sets of fractures cut ductile fabrics associated with tectonic episodes (i) and (ii) outlined above (Alexander and Shail, 1996), and cut and/or displace the stage 3 basalt dykes of Roberts *et al*. (1993). At a larger scale, faults from both sets displace low angle faults related to ophiolite assembly and final obduction (Power *et al*., 1996). It is possible that some of these faults may have initiated in an on ridge setting (e.g. Hopkinson and Roberts , 1995) and have been subsequently reactivated during post-obduction tectonic episodes.

Previous work and field relationships

Fracture-hosted mineralisation in rocks of basic composition (gabbro, Landewednack Schist, Traboe Schist, basalt) exposed of the eastern part of the Lizard peninsula has been well documented (e.g. Seager, 1971). In contrast, little has been published on the fracture-hosted mineralisation in rocks of ultrabasic composition (peridotite). Wall rock composition exerts a fundamental control upon the mineralogy of the fracture infill. Fractures in rocks of basic composition host mineral assemblages dominated by prehnite, calcite, adularia and zeolites (Seager, 1971) whereas fractures in rocks of ultrabasic composition host mineral assemblages dominated by talc, dolomite and sepiolite (this study).

In addition to massive talc, vein talc is present throughout the peridotite. It occurs as a pale green to white, fibrous fracture fill and often shows crack-seal textures. Where talc veins cut basic dykes within the peridotite (and also at the interface between peridotite and acid Kennack Gneiss), saponite is preferentially crystallised. This Mg-smectite is white, has a waxy feel and often has a fibrous habit resembling talc.

Anthophyllite occasionally occurs as a honey coloured fibrous vein fill (e.g. southern Kennack Sands). The fibres may exceed 10 cm in length and are orientated roughly parallel to the vein walls. None of the anthophyllite veins observed cut any other structures, hence the relative chronology cannot be constrained.

Dolomite is the most common fracture-hosted vein mineral within the peridotite. It occurs as white to pink veins which frequently show crack-seal textures; numerous episodes of dolomite formation are apparent. Net veining is particularly well developed in the footwall of the many of the larger fault zones (e.g. the southern end of Pentreath beach [SW 694 126]). Minor amounts of quartz and calcite are associated with the dolomite mineralisation.

Although sepiolite has previously been described from only two localities on the Lizard (Callière and Henin, 1949; Midgley, 1959), it is abundant in fractures within the peridotite. It occurs as fibrous mats which may exceed 3 mm in thickness (e.g. eastern Kennack Sands [SW 741 167]). Individual fibres within these mats are commonly aligned parallel to the direction of fault movement and can be used as kinematic indicators. Sepiolite generally occupies minor fractures that cut previous mineral infills within the main fault zones and is clearly the last formed phase.

Brecciation is commonly associated with mineralisation and mineralised fault zones frequently contain elongate clasts of vein serpentine (Figure 2); vein serpentine clasts are abundant in both talc and dolomite mineralised faults.

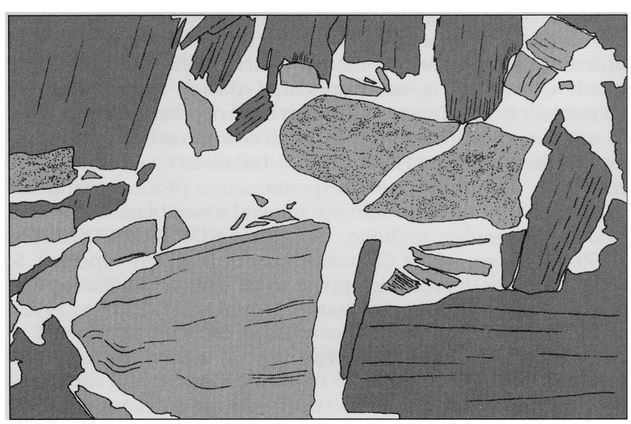

Figure 2. Drawing of a photomicrograph of vein material showing clasts of vein serpentine, serpentinite and talc in a matrix of dolomite. Unshaded areas represent dolomite, dark shading represents talc (showing cleavage), and lighter shading represents serpentinite and vein serpentine; vein serpentine is distinguished from primary serpentinite by its lack of magnetite inclusions and its fine lamination. Field of view is approximately 2.5 mm.

Correlation of fracture-hosted mineralisation between gabbro and peridotite

Direct correlation of fracture-hosted mineralisation between the gabbros and peridotites is difficult due to: (i) the strong dependence of vein mineralogy upon wall rock composition, and (ii) the absence of veins of suitable along-strike persistence at the peridotite-gabbro boundary. Chemical similarities between the gabbro and peridotite vein mineralogy may allow a possible correlation; prehnite ($Ca_2Al[AlSi_3O_{10}](OH)_2$) in the gabbro may be the Ca and Al analogue of the talc ($Mg_3Si_4O_{10}(OH)_2$) formed in the peridotite (Figure 3). Similarly calcite (basic lithologies) seems likely to be equivalent to dolomite (peridotite). Vein orientation, morphology and kinematics also suggest that the mineralisation events may be synchronous (e.g. Alexander and Shail, 1996).

Since carbonate veins occur in both the peridotites and gabbros, stable isotope ratios have been used to "fingerprint" these carbonates in an attempt to correlate fracture controlled mineralisation between the two lithologies. Oxygen and carbon stable isotope ratios were obtained from a number of calcite specimens (mostly sampled from veins within the gabbro) and dolomite specimens (sampled from veins within the peridotite). The samples were dissolved overnight in 100% phosphoric acid at 25°C (100°C for the dolomite samples); the resultant CO_2 was analysed using a mass spectrometer. The results (Table 1) show a distinct, albeit scattered, grouping of both the dolomite and calcite data around a common mean of $\delta^{18}O_{smow}$ 24.29 and $\delta^{13}C_{PDB}$ = -9.96 (Figure 4). Three anomalous data points were excluded and are interpreted as a different, probably earlier, higher temperature (possibly sub-oceanic) event *(cf.* Hopkinson and Roberts, 1995).

Sample No.	Carbonate Phase	Locality	$\delta^{13}C_{PDB}$	$\delta^{18}O_{SMOW}$
5L155	Dolomite	Poltesco	-8.09	23.86
6L01	Dolomite	Kennack Sands	-8.22	24.63
6L04	Dolomite	Kennack Sands	-10.34	22.42
6L05	Dolomite	Kennack Sands	-8.29	26.63
6L06	Dolomite	Kennack Sands	-9.33	23.48
6L07	Dolomite	Kennack Sands	-9.57	28.58
6L10	Dolomite	Pentreath	-11.46	25.53
6L02	Calcite	Kennack Sands	-7.95	26.5
6L03*	Calcite	Dean Quarry	-7.95	13.45
6L12*	Calcite	Dean Quarry	-8.1	13.51
6L13*	Calcite	Dean Quarry	-7.63	12.49
6L14	Calcite	Dean Quarry	-11.3	22.34
6L15	Calcite	Dean Quarry	-11.44	22.62
6L17	Calcite	Dean Quarry	-11.26	22.91
6L18	Calcite	Dean Quarry	-11.43	22.6
6L19	Calcite	Dean Quarry	-11.22	25.96
6L20	Calcite	Dean Quarry	-10.39	23.45
6L21	Calcite	Dean Quarry	-8.36	24.03
6L22	Calcite	Dean Quarry	-10.66	23.06
Mean†			-9.96	24.29

Table 1: Stable isotope results. three anomalous points are excluded from mean and subsequent calculations.

Calcite and dolomite have different oxygen isotope fractionation coefficients, hence calculating a hypothetical fluid oxygen isotopic composition should reveal any isotopic differences between the fluids that precipitated calcite, and those that precipitated dolomite. Assuming fluid temperatures of 200°C and 125°C (respective average fluid temperatures for east-north-east and north-north-west striking structures in south Cornwall; Wilkinson *et al.,* 1995) the oxygen isotopic composition of the fluid can be determined using the equation:

$$(\delta^{18}O_{carbonate} - \delta^{18}O_{fluid}) = \frac{A \times 10^6 + B}{T^2}$$

Where T is temperature (K) and A and B are fractionation coefficients; for calcite A = 2.78 and B = -2.89 (Friedman and O'Neil, 1972); for dolomite A = 3.23, B = -3.29 (Sheppard and Schwarz, 1970). The resulting data (the three anomalous points were omitted) show that again a fair amount of scatter is evident and is unrelated to mineral phase. A slightly smaller standard deviation is apparent at 200°C. The overlapping fields of dolomite and calcite (Figure 4) indicate that the isotopic composition of the mineralising fluids are similar.

DISCUSSION

Timing of Serpentinisation

Cross-cutting relationships provide an initial constraint upon the age of the primary pervasive serpentinisation and the formation of vein serpentine. Vein serpentine post-dates the initial, pervasive serpentinisation episode as rodingites, probably formed during the initial episode of serpentinisation, are cut by the vein serpentine. The fine lamination of the vein serpentine is always preserved (it has not re-crystallised). Furthermore, brecciated clasts of vein serpentine are found in numerous mineralised fault zones and predate fracture-hosted vein mineral phases previously described. This suggests that vein serpentine formation (and therefore the earlier pervasive serpentinisation) pre-dates vein mineralisation.

Ar^{40}-Ar^{39} dating of adularia from similar fracture systems, within basic lithologies, at Porthkerris [SW 805 251] (Seager *et al.,* 1978) and Dean Quarry [SW 805 204] (Seager *et al.,* 1975; Halliday and Mitchell 1976) yields ages of between 210 and 220 Ma; younger ages of 160-170 Ma are interpreted as overprinting by a subsequent hydrothermal event (Seager *et al.,* 1978). Stable isotope ratios (this study) suggest that the mineralising fluids were genetically similar, and that calcite precipitation within the gabbros, and dolomite precipitation within the peridotites were approximately coeval. Hence, the serpentinisation events are older than 210-220 Ma (Triassic), and probably significantly older if talc and carbonate mineralisation are coeval with the eastnorth-east-west-south-west extensional faults (e.g. Alexander and Shail, 1996). Since vein serpentine pre-dates both talc and dolomite mineralisation, it may have formed during the very earliest stages of extensional faulting. It is therefore tentatively suggested that vein serpentine was precipitated in veins associated with the initial stages of north-north-west-south-south-east extension during the latest Carboniferous to early Permian (e.g. Alexander and Shail, 1996). Consequently, pervasive serpentinisation is likely to predate this (i.e. pre latest Carboniferous).

Rodingite alteration affects Kennack Gneiss (e.g. Hall, 1979) which is associated with the early stages of obduction (e.g. Sandeman, 1988). Radiometric dating of the Kennack Gneiss (RbSr isochron; Styles and Rundle, 1984) gives an age of 369 ± 12 Ma. Hence pervasive serpentinisation is therefore likely to have occurred within the continental realm (i.e. post 369 Ma; late Devonian), although the initiation of volumetrically minor localised serpentinisation, within the oceanic realm cannot be discounted.

Vein serpentine is always paragenetically early (Figure 3). Fluids within the Lizard complex therefore do not appear to have been saturated with respect to serpentine group minerals since the precipitation of vein serpentine, and this probably indicates that significant pervasive serpentinisation had essentially halted by this time. This is perhaps surprising considering that the Lizard complex is likely to have been at a high level in the continental crust since the Carboniferous and subject to circulating groundwaters. A number of factors can retard serpentinisation.

As serpentinite acts as a sink for water (e.g. Flett, 1946), a low activity of water may lead to water being removed from the serpentinisation front, temporarily stopping serpentinisation (O'Hanley, 1996). Temperature also controls the rate of serpentinisation; at temperatures around 20°C serpentinisation reactions proceed slowly (e.g. O'Hanley, 1996). As the Lizard complex is likely to have been at a high level within the continental crust, a combination of low temperature and low activity of water seems the most plausible explanation for the lack of significant serpentinisation since the main pervasive episode.

Temperature of serpentinisation

The lizardite-chrysotile assemblage of the primary serpentinisation episode suggests a relatively low temperature (less than 250°C; e.g. O'Hanley, 1996) however, as noted by Floyd *et al.* (1993), this may indicate recrystallisation in a retrograde pressure - temperature regime. Antigorite is present at a few localities (Green, 1964a) and suggests that initial serpentinisation may have taken place at relatively high temperatures (within the antigorite stability field, e.g. 250-400°C; O'Hanley, 1996), perhaps during the early stages of obduction. Re-equilibration to an assemblage of lizardite and chrysotile may have occurred as the peridotite was exhumed. Lizardite may pseudomorph early antigorite flame structures giving rise to the coarse lizardite described by Floyd *et al*. (1993). Similarly, magnetite veins are abundant within the rims of kernel structures and are indicative of recrystallisation of serpentinite (e.g. O'Hanley, 1996). The lizardite-chrysotile assemblage of the vein serpentine suggests a formation temperature of less than 250°C.

Nature of serpentinisation

Kernel structures, present throughout the peridotite, indicate that mass may have been conserved during pervasive serpentinisation; rodingites are present but are volumetrically insignificant. An volumetric expansion of the peridotite may therefore have accompanied serpentinisation.

Massive talc only occurs in close proximity to acid Kennack Gneiss. Flett (1946) proposed that the talc at Gew Graze formed within the gneiss, rather than the peridotite. This led Hall (1979) to suggest that the talc may form during intense Mg metasomatism associated with the serpentinisation process. However this locality is no longer exposed (and has not been for over 100 years; e.g. Flett, 1946) and elsewhere on the Lizard (e.g. the north end of Pentreath beach [SW 692 129]) talc occurs as pods and screens within the acid Kennack Gneiss. The relict structures seen in the massive talc suggest affinities with peridotite rather than the gneiss. Massive talc may therefore be a product of Si metasomatism of peridotite enclosed by, or in close proximity to, acid gneiss. Where xenoliths are enclosed in more mafic fractions of Kennack Gneiss, a small reaction rim (< 10 mm) of actinolite, chlorite and talc is developed. This assemblage is typical of rodingite alteration (e.g. Coleman, 1977) and suggests that the silica content of the enclosing rock controls the extent and type of the metasomatism. Rodingite mineral assemblages of prehnite and amphibole described by Hall (1979) from basaltic dykes within the peridotite may be a result of the regional prehnite-pumpellyite grade metamorphism (e.g. Barnes and Andrews, 1981) rather than Ca metasomatism during serpentinisation.

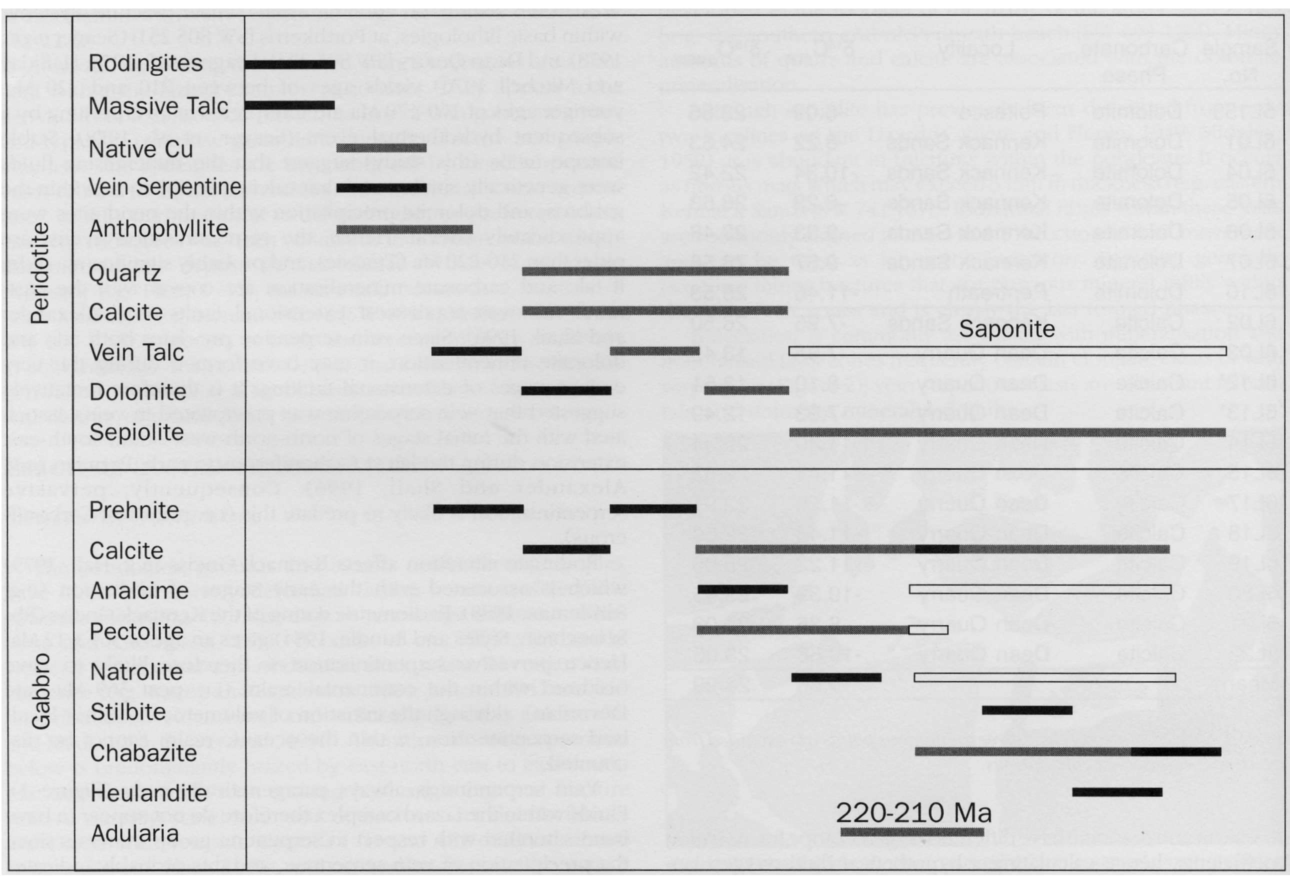

Figure 3: Paragenetic sequence for the mineralisation across the Lizard complex. Solid boxes represent crystallisation events; shaded boxes represent uncertainties regarding the exact placing of an event; unfilled boxes indicate recrystallisation to clays (typically to saponite or montmorillonite clays). Paragenetic sequence within the gabbro units is modified after Seager (1971; 1978).

Fracture-hosted mineralisation

Prehnite in the gabbro can be tentatively linked with talc in the peridotite, and calcite within the gabbro can be correlated with dolomite within the peridotite (Figure 3). Inconsistent crosscutting relationships between talc and dolomite are similar to relationships between prehnite and calcite in the gabbros (e.g. Seager, 1971). Hence, by analogy with the gabbro mineralisation, two phases of talc formation separated by an episode of dolomite formation are proposed. Adularia is present in the peridotite but is hosted by basic dykes and indicates that wall rock chemistry exerts a profound influence on the vein mineralogy even at a centimetre scale. Placing the formation of anthophyllite within the paragenetic sequence is problematic as no cross cutting relationships have been observed. However, as anthophyllite typically forms at higher temperatures than talc (e.g. Hemley *et al.,* 1977; O'Hanley, 1996) it may pre-date talc mineralisation.

Sepiolite typically forms from saline, high pH, moderately siliceous fluids (e.g. Wollast *et al.,* 1968). Chemically similar saline brines, derived from Permo-Triassic basins, are associated with north-north-west striking crosscourse mineralisation (e.g. Scrivener *et al.,* 1986; Scrivener *et al., 1994).* It is therefore tentatively suggested that sepiolite precipitation in the peridotite may be related to crosscourse mineralisation elsewhere. Similarities between the morphology of saponite and talc suggest that saponite may pseudomorph talc during a subsequent event, perhaps related to the formation of sepiolite.

Mineralisation within the Lizard complex can be tentatively correlated with the regional tectonic regime (Alexander and Shail, 1996). Talc and dolomite in the peridotite, and prehnite and calcite in the gabbro may be synchronous with north-north-west—south-south-east extension (Alexander and Shail, 1996) which has been dated at approximately 300-260 Ma (e.g. Wilkinson *et al.,* 1995). Sepiolite in the peridotite, and zeolites and adularia in the gabbro may be coeval with late Permian strike-slip faulting and Triassic east-north-east—west-south-west extension (Alexander and Shail, 1996); mineralisation associated with these events has been dated at 270-230 Ma (e.g. Wilkinson, 1995).

Although carbonate mineralisation is widespread across the complex, dolomite rather than magnesite the dominant carbonate phase within the peridotites. The absence of magnesite may be due to serpentinisation pre-dating the carbonate mineralisation, thereby "locking" up Mg ions and lowering the activity of Mg, leading to the preferential precipitation of dolomite.

CONCLUSIONS

At least two episodes of serpentinisation of the Lizard peridotite are observed. The first episode is responsible for the pervasive alteration of the peridotite to lizardite, chrysotile and magnetite. The second (later) episode gives rise to the precipitation of lizardite and chrysotile vein serpentine. A volume increase accompanied pervasive serpentinisation giving rise to expansion-related kernel structures. Pervasive serpentinisation may have initiated during obduction at temperatures of 300-400°C with the formation of antigorite, and subsequently recrystallised to the lower temperature lizardite and chrysotile assemblage as the peridotite was exhumed.

Host rock chemistry exerts a strong control on the mineralogy of the vein systems on the Lizard. Calcite-prehnite-zeolite mineralisation is restricted to basic units whilst dolomite-talc-sepiolite mineralisation is characteristic of the ultrabasic units. Stable isotope ratios indicate that the fluids responsible for fracture-hosted mineralisation within the gabbros and peridotites are genetically similar, and probably coeval. Previously published radiometric ages, from the gabbros, can

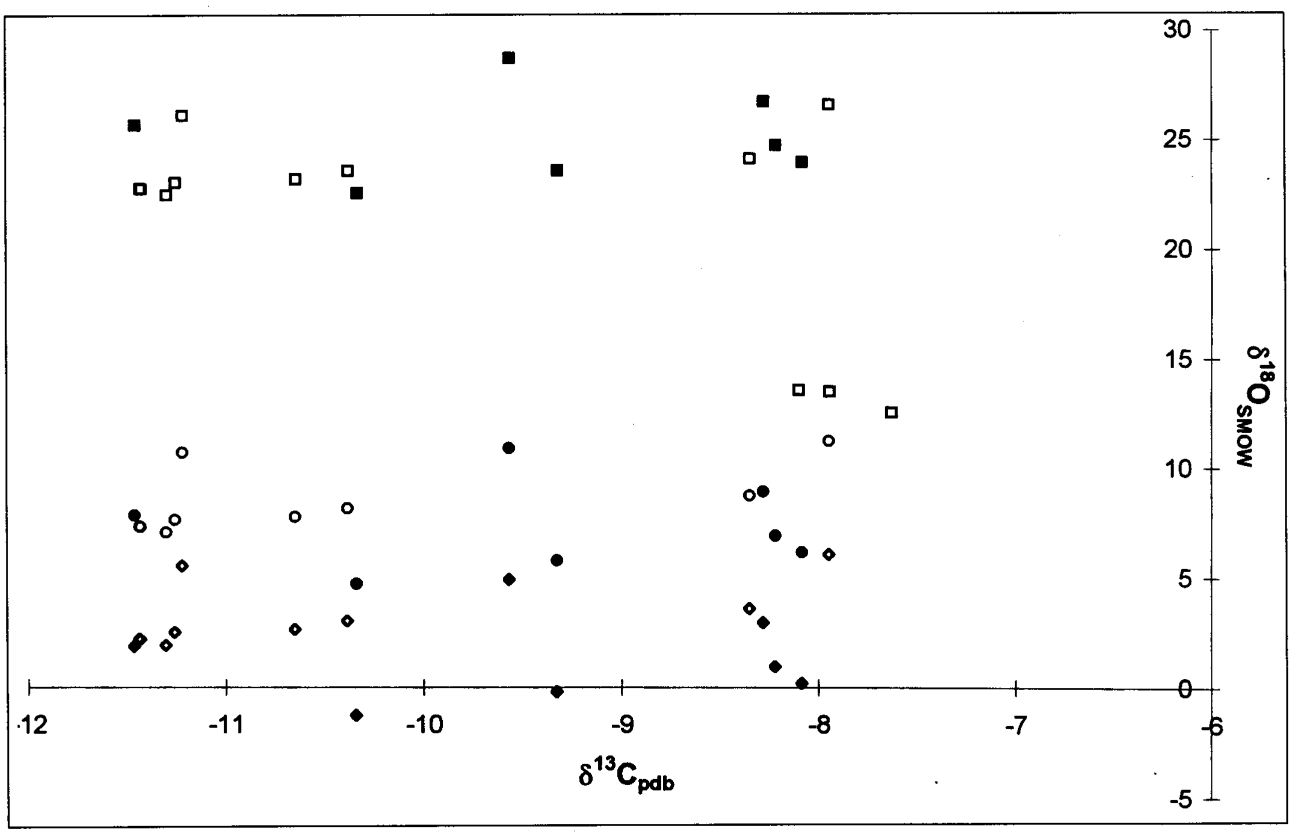

Figure 4: Stable isotope results (calcite is represented by unfilled symbols, dolomite by filled symbols). Squares represent raw data. Diamonds and circles represent calculated hypothetical mineralising fluid isotopic compositions; calculated using a hypothetical fluid temperature of 125°C (diamonds) and 200°C (circles).

therefore be applied to the fracture-hosted mineralisation within the peridotite. Vein serpentine predates all mineralisation and is interpreted as synchronous with the earliest extensional faulting and therefore latest Carboniferous to early Permian in age.

Pervasive serpentinisation appears to predate vein serpentine formation, hence pervasive serpentinisation may have effectively ceased by the late Carboniferous to early Permian (although minor, localised serpentinisation may have occurred for some time after); the protracted weathering hypothesis (e.g. Flett, 1946) appears inappropriate. As pervasive serpentinisation affects obduction related lithologies (e.g. the Kennack Gneiss), it is suggested that serpentinisation occurred after the initiation of obduction (i.e. post late Devonian). However, a pre-obduction initiation of the serpentinisation process cannot be discounted. In conclusion, a late Devonian to Carboniferous age is suggested for the pervasive serpentinisation of the Lizard peridotites.

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