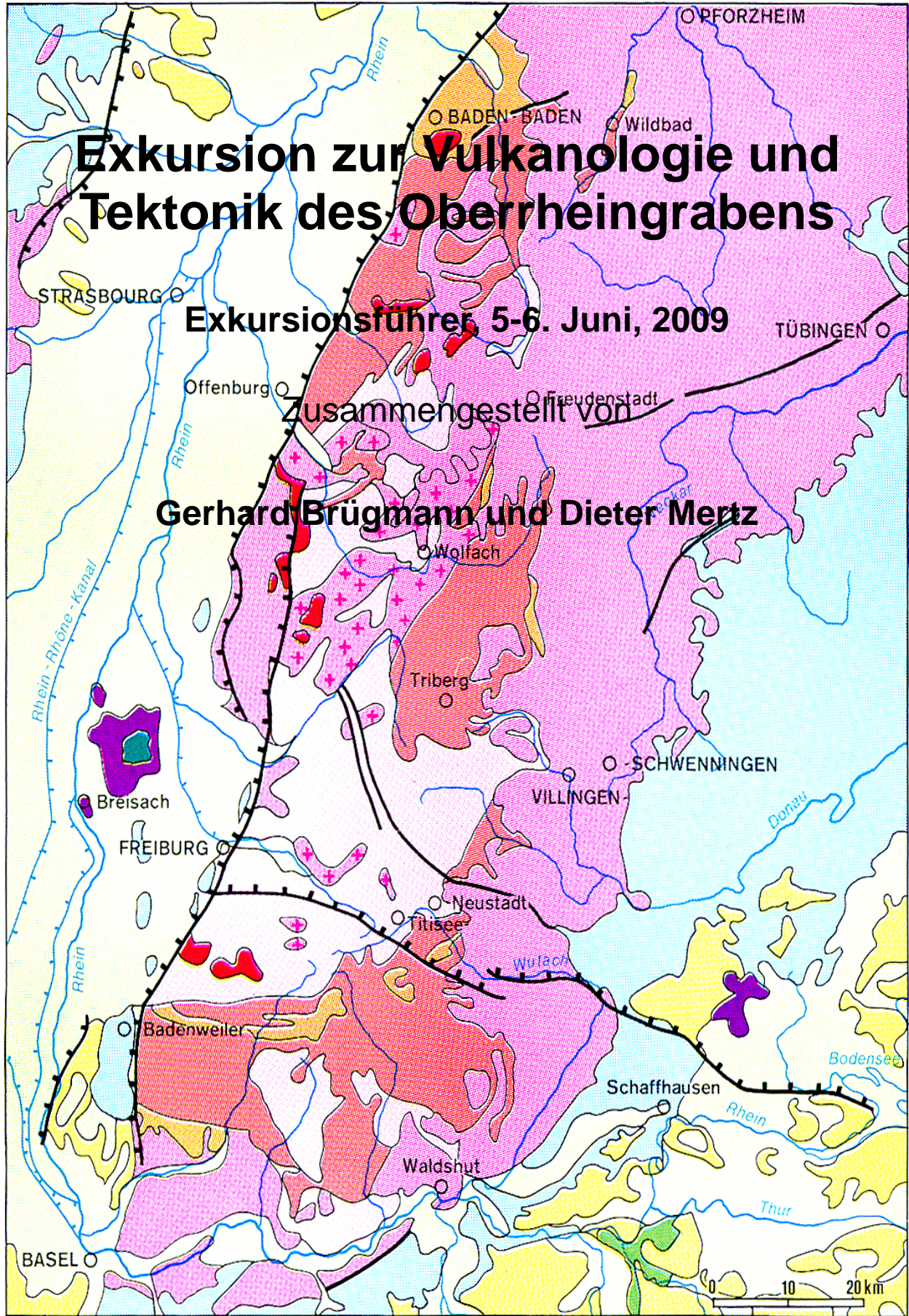


Exkursion zur Vulkanologie und Tektonik des Oberrheingrabens

Exkursionsführer 5-6. Juni, 2009

Zusammengestellt von

Gerhard Brüggemann und Dieter Mertz

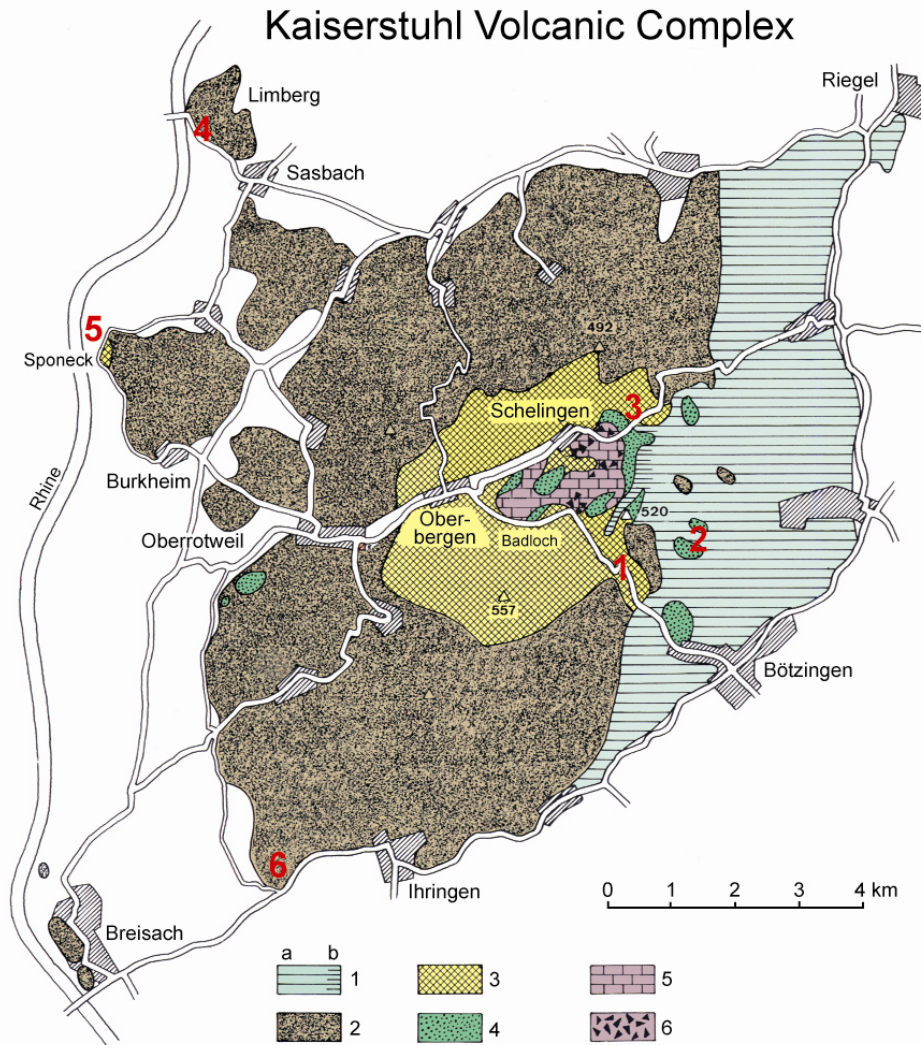


Field Trip Schedule

05. Juni 2009 *Geologie des Kaiserstuhl-Vulkankomplexes*

07:00 Busabfahrt, vor dem Geo-Institut

9:30 Ankunft Vogelsangpass zwischen Oberschaffhausen und Vogtsburg



- 1a) Oligocene to Jurassic Sediments
- 2) Lava flows and pyroclastic rocks
- 4) Phonolite
- 6) Subvolcanic breccia

- 1b) Contact metamorphic Oligocene
- 3) Subvolcanic Essexite rocks
- 5) Carbonatite
- white: Quaternary deposits;
- Loess cover not shown

1. Vogelsangpass, Einführung Oberrheingraben und Kaiserstuhlüberblick
2. Phonolit bei Endhahlen, SWW von Eichstätten
3. Karbonatit und subvulkanische Breckzie am Ohrberg, Schelingen
4. Limburgite und Nephelinite am Limberg und Lützelberg, NW Sasbach
5. Essexit-Intrusion und Tephrit-Laven am Fuß von Kastell Sponeck
6. Tephritlaven am Vulkanfelsen, westlich Ihringen

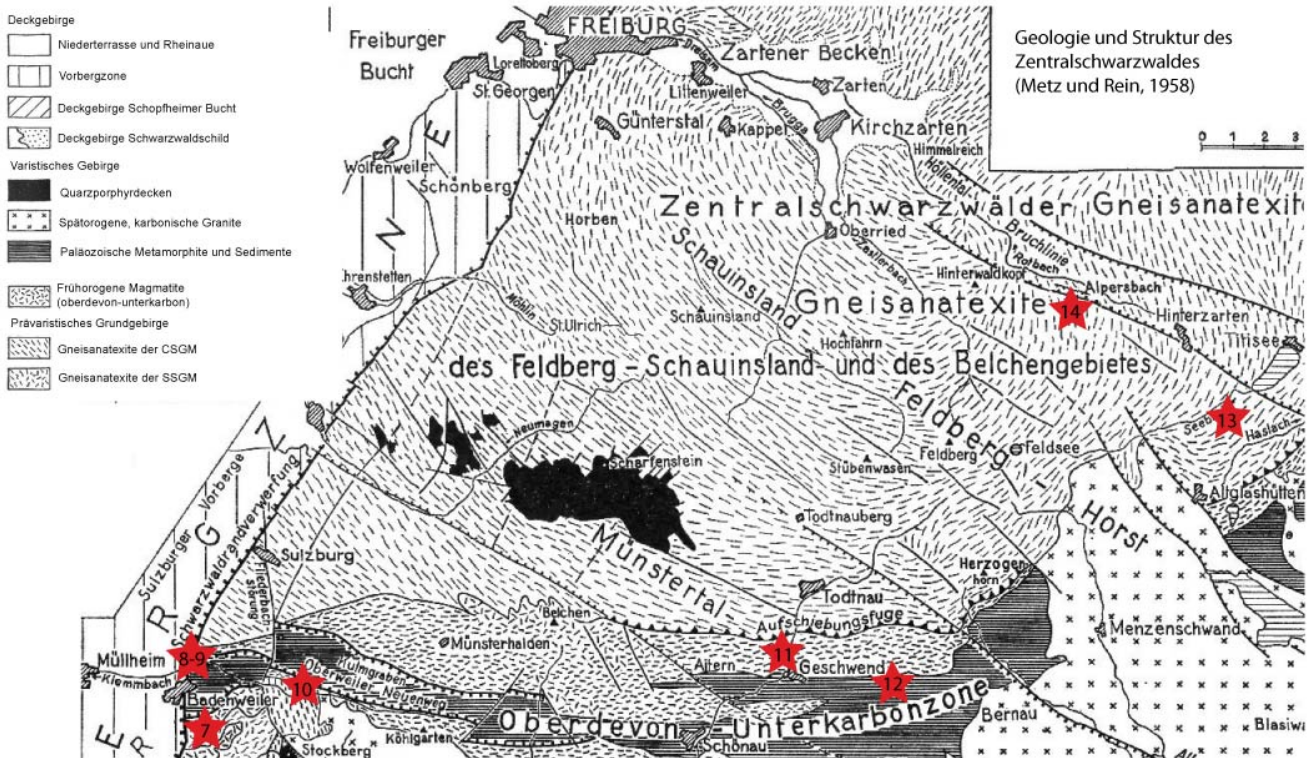
19:00 Uhr Unterkunft Gast zum Löwen in Sasbach

Field Trip Schedule

06. Juni 2009 *Geologie Rheingrabengrand und südlicher Schwarzwald*

7:30 Uhr Frühstück

8:00 Uhr Abfahrt Richtung Badenweiler



7. Quarzfing, S Badenweiler, Wanderung Klinik Hausbaden – Alter Mann
8. Schwärzestraße, nördlich Badenweiler, Rheingrabenrandstörung (Muschelkalk/Kulmkonglomerat)
9. Parkplatz Schwärze, Wanderung entlang Geopfad: Lias, Dogger, Tertiär,
10. Weiterfahrt Landstraße L131 Richtung Schönau, Malsburg-Granit
11. Schönau-Geschwend, Randgranit
12. Wanderung Sengalenkopfweg, Randgranit-Sengalenkopfschiefer-
13. Stierhütte / Falkau, Migmatitischer Paragneis der Unit 1 im CSGC
14. Lochütte/Alpesbach, WSW von Hinterzarten; Eklogite der Unit-1 im CSGC

Rückfahrt über Kirchzarten - Freiburg – A5 - Mainz

A. The Upper Rhine Graben and the Kaiserstuhl Volcanic Complex

Introduction

Two geodynamic events affected the West-European plate during the Cenozoic:

- (1) the Alpine orogeny, which is related to convergence between the African and European plates, and which started during the Late Cretaceous and is still ongoing;
- (2) the opening of the European Cenozoic Rift System (ECRIS) that started during the Late Eocene (Fig. 1).

The question of whether and how the Alpine orogeny and the opening of the ECRIS are related to each other are still debated. A variety of hypotheses for its origin have been proposed, involving processes such as plume-related active rifting, foreland splitting in response to the Alpine orogeny, back-arc rifting or slab-pull. While the Alpine orogeny produced crustal thickening in the Alps and in the Pyrenees, it also entailed long-wavelength lithospheric folding of the European foreland (Dèzes et al., 2004). Thermal erosion of the lithosphere by up-welling asthenospheric thermal anomalies may also have played a role in the long-wavelength deformation of the European lithosphere.

The opening of the ECRIS produced lithospheric thinning in a number of grabens extending over a distance of some 1,100 km from the shores of the North Sea into the Western Mediterranean, and from the North Sea to the North-West to the Bohemian Massif to the East (Fig. 1)

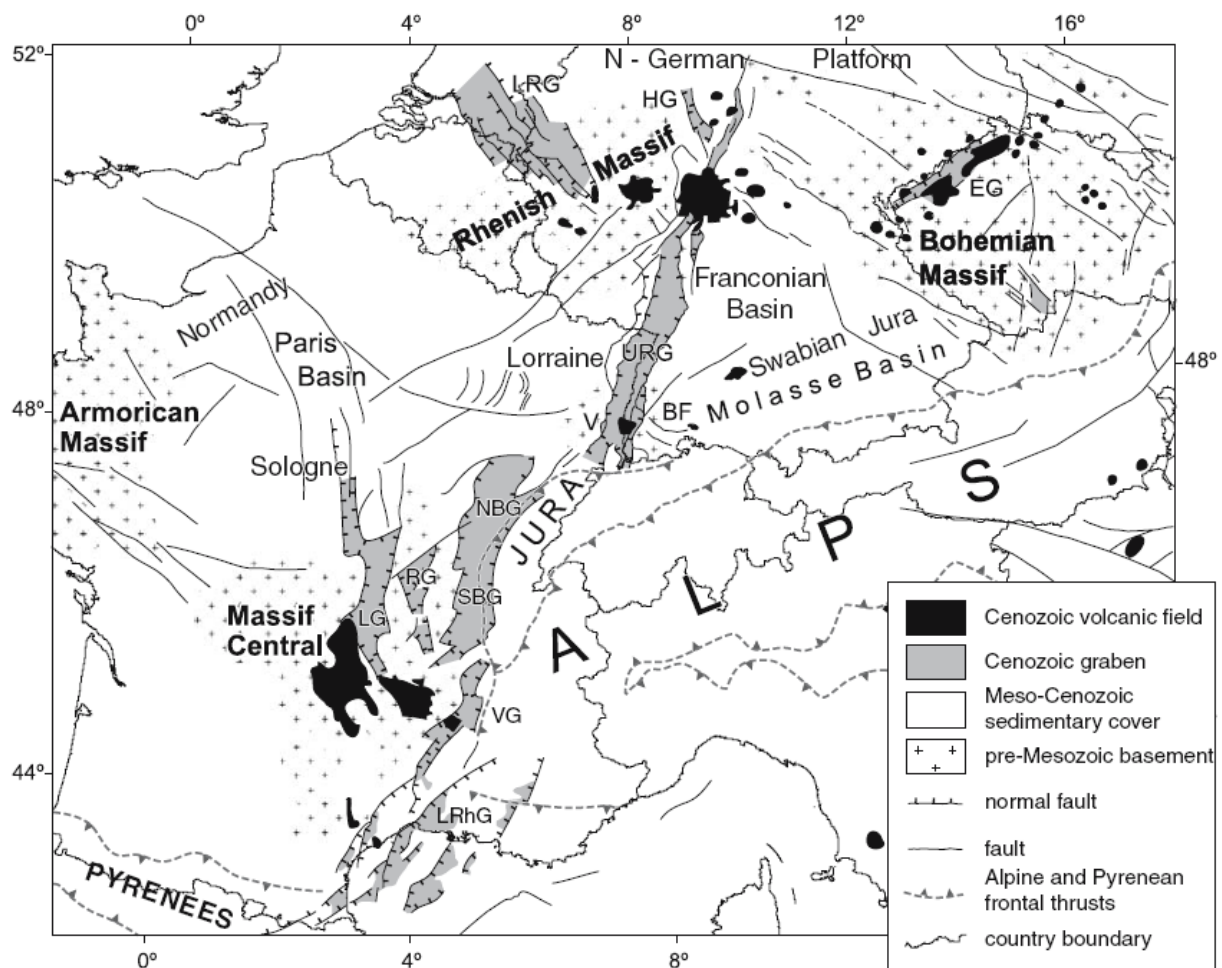


Fig. 1: Structural map of the ECRIS in the NW-Alpine foreland (after Bourgeois et al. 2007)

BF Black Forest, EG Eger (Ore) Graben, FP Franconian Platform, HG Hessian grabens, LG Limagne Graben, LRG Lower Rhine (Roer Valley) Graben, LRhG Lower Rhone Grabens, NBG Northern Bresse Graben, RG Roanne Graben, SBG Southern Bresse Graben, URG Upper Rhine Graben, V Vosges, VG Valence Graben

Anorogenic intra-plate magmatism was widespread in Europe from early Tertiary to recent times, extending west to east from Spain to Bulgaria, and south to north from Sicily to central Germany (Wilson and Downes, 2006). Magmatism is spatially and temporally associated with Alpine-Pyrenean collisional tectonics, the development of an extensive lithospheric rift system in the northern foreland of the Alps (ECRIS), and, locally, with uplift of Variscan basement massifs (Massif Central, Rhenish Massif, Bohemian Massif; Fig. 1). The volcanic regions vary in volume from large central volcanoes (e.g. Cantal, Massif Central; Vogelsberg, in central Germany), to small isolated plugs (e.g. Kaiserstuhl volcanic complex, Urach and Hegau provinces in southern Germany). The volcanic rocks are predominantly sodic (melilitites, nephelinites, basanites and alkali olivine basalts); however, locally, potassic magma types (olivine leucitites, leucite nephelinites) also occur. In several localities (e.g. Vogelsberg and the Rhine Graben, Germany) olivine and quartz tholeiites form a significant component of the magmatism (Wedepohl and Baumann, 1999). The sodic magmas were derived by variable degrees of partial melting (ca. 0.5-5%) within a transitional zone between garnet-peridotite and spinel-peridotite mantle facies, close to the base of the lithosphere. The potassic magma types are interpreted as partial melts of enriched domains within the lithospheric mantle. Mantle partial melting was induced by adiabatic decompression of the asthenosphere, locally in small-scale, plume-like, diapirs, which appear to upwell from ca. 400 km depth (Wilson and Downes, 2006).

Evolution of the Upper Rhine Graben

The Upper Rhine Graben (URG) forms the central part of the Cenozoic Central European Rift System Graben (Fig. 1). Limited to the north by the Rhenish Massif and the Swiss Jura to the south, the NNE trending 30 to 40 km wide graben runs over 300 km from Basel to Frankfurt crossing boundaries of various tectono-stratigraphic units of the Hercynian internides. Figure 2 (Sissingh, 2003) gives a structural overview indicating the Upper Rhine Graben to bifurcate into the northwestern Lower Rhine Graben by crossing the Hercynian Rhenish Massif, and the northern Hessian Depression, respectively. From a structural point of view, the evolution of this northwestern branch within the Rhenish Massif (“Middle Rhine Graben“) is not well understood, because the slaty Devonian-Carboniferous low-grade metamorphic sequences appear to have behaved ductile in response to extension which could have obscured much of the physiography of the graben structure. Structural and geological arguments indicate that the localization of the Cenozoic Rift System is presumably controlled by pre-existing zones of crustal weakness of Hercynian or possibly pre-Hercynian origin (Schumacher, 2004).

In the north, the Upper Rhine Graben is separated from the Lower Rhine Graben and Hessian Depression rift segments, respectively, by an E-striking belt of intra-plate volcanic fields (Eifel, Siebengebirge, Westerwald, Vogelsberg, Rhön) of Eocene to Quaternary age (Fig. 2). In the south, it is bordered by the “Jura Molasse“ (Fig. 1), representing terrestrial or shallow marine sediments deposited in the Alpine foreland basin during the Tertiary Alpine orogeny. To the east and west, the Upper Rhine Graben is framed by Hercynian and pre-Hercynian basement complexes composed of plutonic, volcanic or metamorphic rocks, which are completely (Pfälzer Wald) or partly (Odenwald, Vosges, Black Forest) covered by mainly Triassic sediments.

In the URG intense exploration for hydrocarbons, mining of potash salts and production of geothermal energy have yielded a large set of geophysical data that has become accessible during the last decade. Depth contours of the Moho (e.g. Ziegler and Dezes, 2005; Sissingh, 2003) show a reduction of crustal thicknesses from 30 km at neighboring regions towards the URG rift zone. The depth of the crust-mantle boundary increases gradually from 24 km in the southern rift segment beneath the site of the Kaiserstuhl volcanic complex, to 26.5 km in the central parts around Karlsruhe and, finally, to 28 km at the northern termination of the URG. The higher Moho uplift in the southern URG may be attributed to an interaction of rift-related thinning and lithospheric bulging in the northern foreland of the Alps. This in turn corresponds to the generally SE-NW directed convergence and coupled collision of the European Plate and the Apulian Microplate during the anti-clockwise convergence of Africa and cratonic Europe.

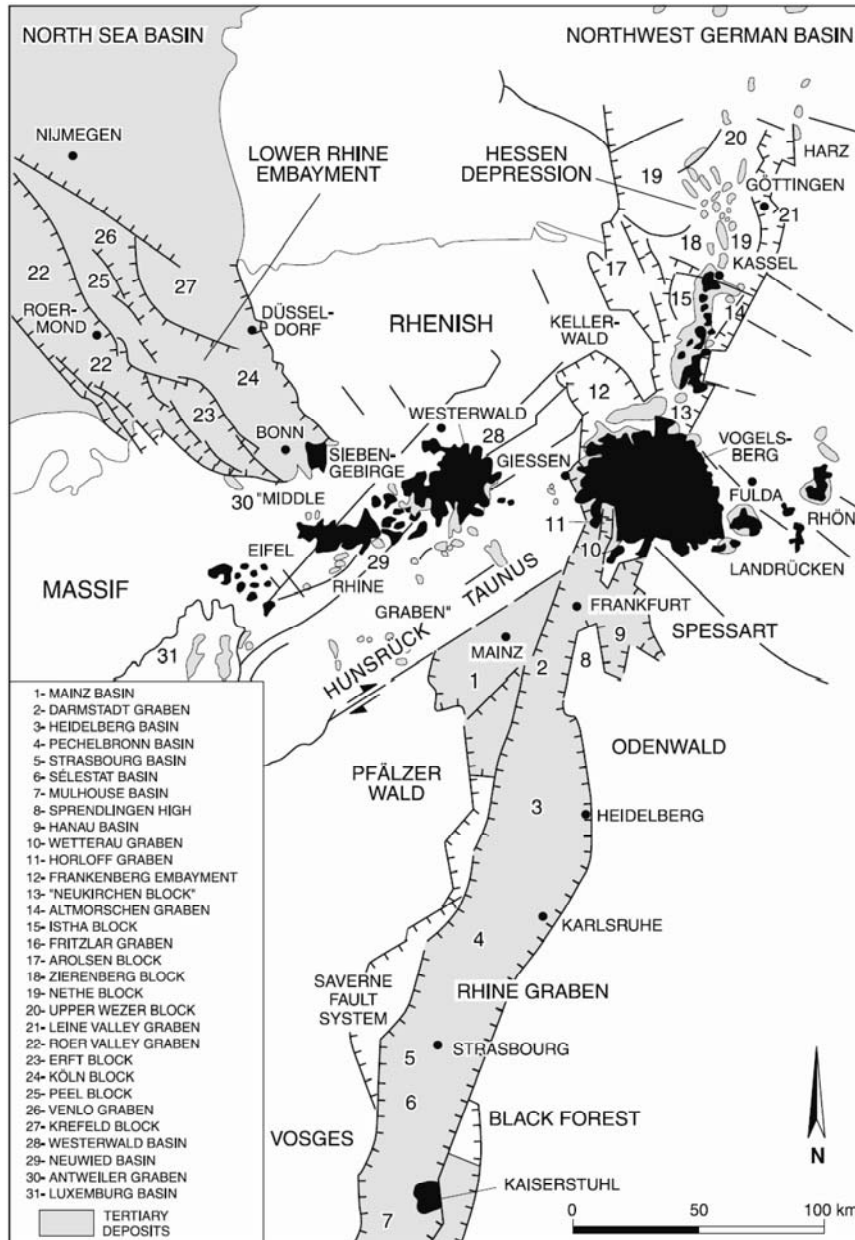


Fig. 2: Structural summary map of the URG in Central and southern Germany (after Sissingh, 2003)

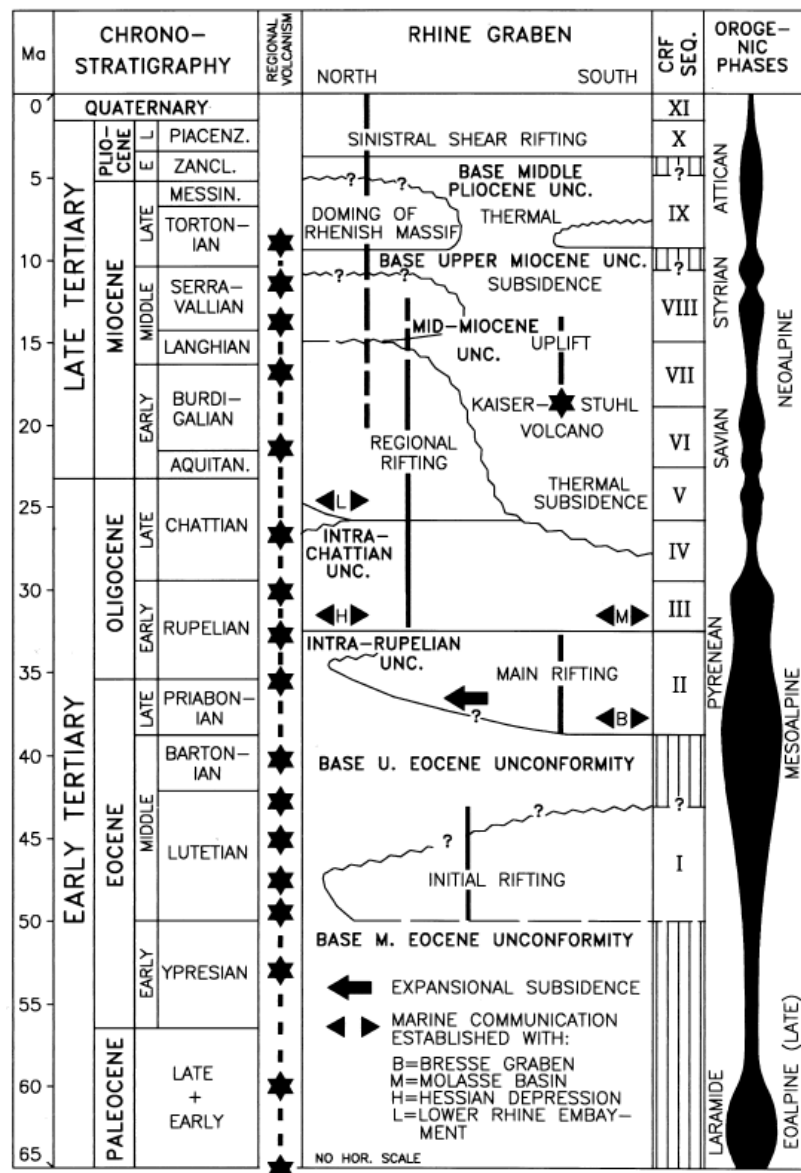
The URG has developed in a region with a complex pre-Tertiary crustal architecture. This had significant influence on the graben geometry (Schumacher, 2004). The NNE-SSW trend of the URG is the so-called Rhenish trend, which was established during the Paleozoic Hercynian Orogeny. Two other Hercynian trends (Erzgebirgian: NE to ENE; Hercynian: NW) are also present. The Hercynian trends are founded in the basement which consists of a sequence of NE trending highs and troughs separated by fault zones. Therefore, the Upper Rhine Graben is segmented into sub-basins and sills, which obliquely cross the graben axis. These ENE oriented structures suggest that the Rhine Graben developed a segmented pull-apart system when the original about N-S compression was replaced by SE-NW compression associated with block movements parallel to oblique strike-slip fault zones during relatively slow subsidence (e.g., Charowicz and Deffontaines, 1993). The total horizontal extension normal to the graben axis amounts to ca. 6 km. While it displays remarkable horizontal east-west symmetry, this is not the case vertically; the depth to the crystalline basement is greater in the east than in the west of the graben. This asymmetry is also reflected by the elevation of the graben shoulders that are, on average, 200 m higher in the east than in the west. The maximum vertical offset between the elevated graben shoulders and the basement amounts to 4500 m. The western shoulder is displaced by about 30 to 40 km towards the south relative to the eastern flank. The graben is still

seismically active and lateral movements between graben and shoulders of about 0.5 mm per year have been observed.

Within the URG Mesozoic sediments rest discordantly on Permo-Carboniferous series and the basement. In the northern URG, Mesozoic sediments comprise Triassic units while in the southern URG the Mesozoic also comprises Jurassic sediments. Over the entire length of the graben, the Cenozoic graben fill in turn overlies these Mesozoic sediments discordantly. It reaches its maximum thickness of syn-rift, post-cretaceous sediments with close to 3400 m in its northern part (e.g. Berger et al., 2005a, b).

Figure 3 (Sissingh, 1998) relates sediment stratigraphy to structural and volcanic events during the Rhine Graben formation. Rift-induced subsidence started close to the Early (Ypresian) - Middle (Lutetian) Eocene transition probably along NNE oriented strike-slip fault zones. Although until the Early (Rupelian) - Late (Chattian) Oligocene transition at ca. 30 Ma mainly the southern Rhine Graben was effected by subsidence, the restricted lacustrine sedimentation of the Messel formation in the northern trough is interpreted to represent first indications of subsidence starting at ca. 48 Ma (Mertz & Renne 2007).

After a period of relatively quiescence in the late Middle Eocene (Bartonian), major rifting took place during the late Eocene (Priabonian) to Early Oligocene resulting in an Early Oligocene transgression from the south. Eventually, in the Middle Oligocene a continuous strait formed connecting the southern subalpine Molasse fore deep with the Hessian Depression which was ultimately fed from North Atlantic water. Multiple halite-dominated salt formation up to ca. 300 m thick (e.g., “Untere



Salzabfolge“) indicate repeated closure of the Rhine Graben basin during Bartonian to Middle Oligocene times. Extension continued until the Early Miocene but was strongly reduced during Late Miocene times. This is presumably related to a major change in stress conditions (Schwarz and Henk, 2005). Starting at around 30 Ma, major subsidence shifted to the northern part of the trough, however, regional changes in facies and sedimentation thicknesses indicate multiple relocation of the trough axis. After this major shift, a last phase of marine (to brackish) sedimentation took place in the northern through indicating transgression from the Hessian Depression at the Oligocene/Miocene transition.

Fig. 3: Stratigraphic diagram showing the main structural events of the Rhine Graben and the volcanic events in and near the Rhine Graben. After Sissingh (1988)

Since the Late Pliocene (Piacenzian) the Rhine Graben has been affected by sinistral shear and transtensional deformation (Fig. 3). Recent dynamic slip on graben faults and neotectonic activity is confirmed by geodetic measurements. This seismic activity is suggested to be due to ongoing subduction of the European continental lithosphere into a still active Alpine orogen causing compressional deformation in the northern Alpine foreland. In addition, there are substantial differences in sediment thicknesses of Late Pliocene and Quaternary fluvial graben sediments with ca. 1 km sedimentation at Heidelberg (“Heidelberger Loch“) representing the most extreme site of sink. Differing sediment thicknesses indicate regional-scale differences in subsidence and on no account graben-forming processes are terminated.

As indicated in Figure 3, volcanism seems to have accompanied the Rhine Graben formation more or less continuously from pre-rifting times (pre-Lutetian) until the Tortonian (e.g. Lippolt, 1982; Keller, 2002). However, many of the data compiled in Figure 3 represent conventional K/Ar ages measured on whole rock samples, which are known to suffer from secondary remobilization, very often from either Ar loss or excess Ar. Therefore, the available age data may give an overview of the geochronological evolution of the volcanism, however, one must appreciate that the detailed interaction between volcanic activity and rifting is not well established.

There are several oil and gas fields in the Upper Rhine Graben mainly related to lower Tertiary or Mesozoic sediments. They occur either in sand deposit facies traps or in tilted blocks where hydrocarbon migration is limited by sealed faults. Diametrical hydrocarbon migration caused by the general eastern dip of the graben sediments explains why the important gas and oil fields (e.g. Landau/Palatina, Pechelbronn/Alsace) are mainly located in the western part of the graben. At various sites, the Rhine Graben provides suitable conditions for geothermal energy production. As an example, at Soultz-sous-Forets (Alsace) a hydrothermal reservoir in granitic basement rocks at a depth of ca. 5 km is being developed delivering about 200°C hot water. There are only a few significant hydrothermal mineralizations genetically related to the Rhine Graben structure (e.g., “Badenweiler Quarzriff“) although the geological setting is highly promising for triggering ore forming processes. Thus, the Rhine Graben is bordered by prominent eastern and western fault systems which provide open tectonic pathways for percolating hydrothermal solution. It is also characterized by a high temperature gradient as well as volcanism-related melt production.

Evolution of the Kaiserstuhl Volcanic Complex in the Southern URG

Since more than 200 year the Kaiserstuhl volcanic complex is the subject of geological and mineralogical studies. The first geological map has been published in 1957. More recently a new map is available, because during the late 20th century spacious reshaping of the landscape to meet new agricultural demands significantly changed the morphology and therefore the geological outcrop conditions (Willmanns et al., 1989; Wimmenauer, 2003). The name of the complex is not derived from the armchair-like shape of the complex. Rather it commemorates the place of an open-air tribunal of the Karolinger Emperor Otto III. in 994.

The Kaiserstuhl is located about 20 km NW of Freiburg (Fig. 2, 4). It is an island in the southern URG with a maximum elevation of about 270 m above the surrounding alluvial plain of the River Rhine (Totenkopf, 557 m). It has a roughly elliptical outline with axes of 16 and 12 km and covering an area of 92 km². The backbone of the Kaiserstuhl is a horseshoe-shaped ridge consisting of a sequence of pyroclastic rocks and lava flows which encloses a basin with intrusive rocks, including the carbonatite at the Badberg massif (Badloch at location 2 in Fig. 5). What we see today are the last ruins of a strato-volcano, which was several hundreds of meter higher than today and surrounded by several smaller parasitic volcanoes, such as the Limberg (Sasbach) or the Humberg (Burkheim). After the volcanic activity ceased about 16 Ma ago, the volcano suffered erosion for several millions years until it has been enclosed by Pleistocene loess sediments. Today, up to 30 m thick loess covers three quarters of the mostly anthropogenically shaped surface (grape-vine terraces). The complex is surrounded by unconsolidated rocks, which have been deposited by the River Rhine and its tributary streams during the Pleistocene and Holocene.

The Kaiserstuhl is a Miocene alkaline volcanic complex belonging to the larger Tertiary eruptive province of Central Europe. It is a unique feature because it is the only large volcano within the Rhine Graben and because of the occurrence of carbonatite rocks. The prominent geophysical feature of the area, the rise of the Moho to only 24 km depth right underneath the Kaiserstuhl is interpreted as

expression of a diapiric mantle dome centered beneath the volcanic complex (Fig. 4).

The volcanic activity in the Kaiserstuhl lasted according to recent data from about 18 to 16 m.y. B.P. (Schleicher and Keller, 1991; Kraml et al., 2006). Small occurrences of dikes and pipes of olivine nephelinite and olivine melilitite on both graben shoulders show a much larger age spread with apparent K-Ar ages from close to 100 - 13 m.y. B.P. These occurrences are concentrated in the area of mantle doming and crustal upwarping. The location of the Kaiserstuhl is likely tectonically controlled, because two major fault system cross beneath the volcano. One is striking along a SE-NW direction reaching the URG coming from the Bodensee area, through the Hegau volcanic province and the Hercynian basement of the Black Forrest. The second one is the SSW-NNO trending URG itself.

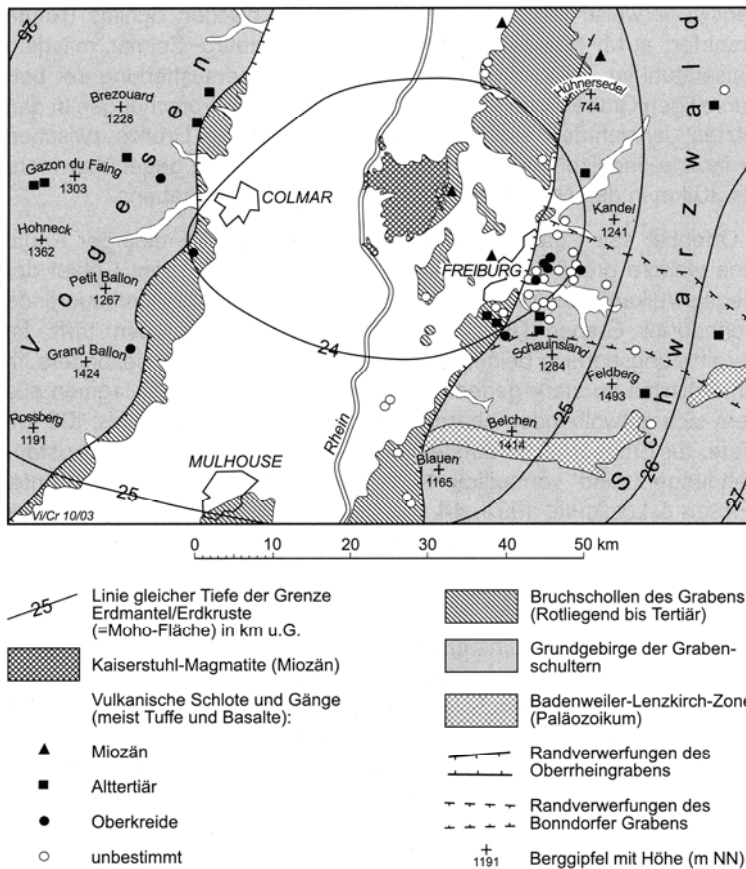


Fig. 4: Regional geology of the Kaiserstuhl volcanic complex. Note the diapiric dome of the Moho beneath the volcanic complex. (After Wimmenauer, 2003)

Major rock types and temporal evolution of the Kaiserstuhl volcanic complex

Two thirds of the igneous rocks outcropping today are lava flows and pyroclastic rocks (Fig. 5). The remaining rock types are subvolcanic breccias and intrusions of silicate and carbonatite melts. Most of the lava flows are less than 10 m thick and can only be traced on the scale of the outcrop. However, some flows, such as the Limburgite flow, can be followed for 1.5 km and can be up to 50 m thick. This lava flow shows locally pahoehoe structures, others represent Aa- or block lavas. Wimmenauer (2003) has described the structure of the volcano and the petrography of the Kaiserstuhl rocks in great detail. The large variety of rock types has been grouped into two petrographic families:

Tephritic-essexitic family

- olivine nephelinite
- limburgite (basanite)
- leucite tephrite
- essexite, theralite
- monzonite,
- mondhaldeite

Phonolitic family

- phonolite
- tinguaite
- haunophyre
- ledmorite
- sodalite syenite
- leucitophyre

Characteristic minerals of the two groups are:

Ti-Augite, plagioclase,

Leucite/analcime or nepheline

Aegirine-bearing augite, melanite

Alkali feldspar, sodalite-hauyne

Titanite

However, in some cases tephritic rocks contain alkali feldspar as well as a mineral of the sodalite-group. In turn the phonolitic rocks occasionally have leucite, analcime, nepheline or plagioclase. In some cases hornblende, biotite, olivine, perovskite and wollastonite have been observed.

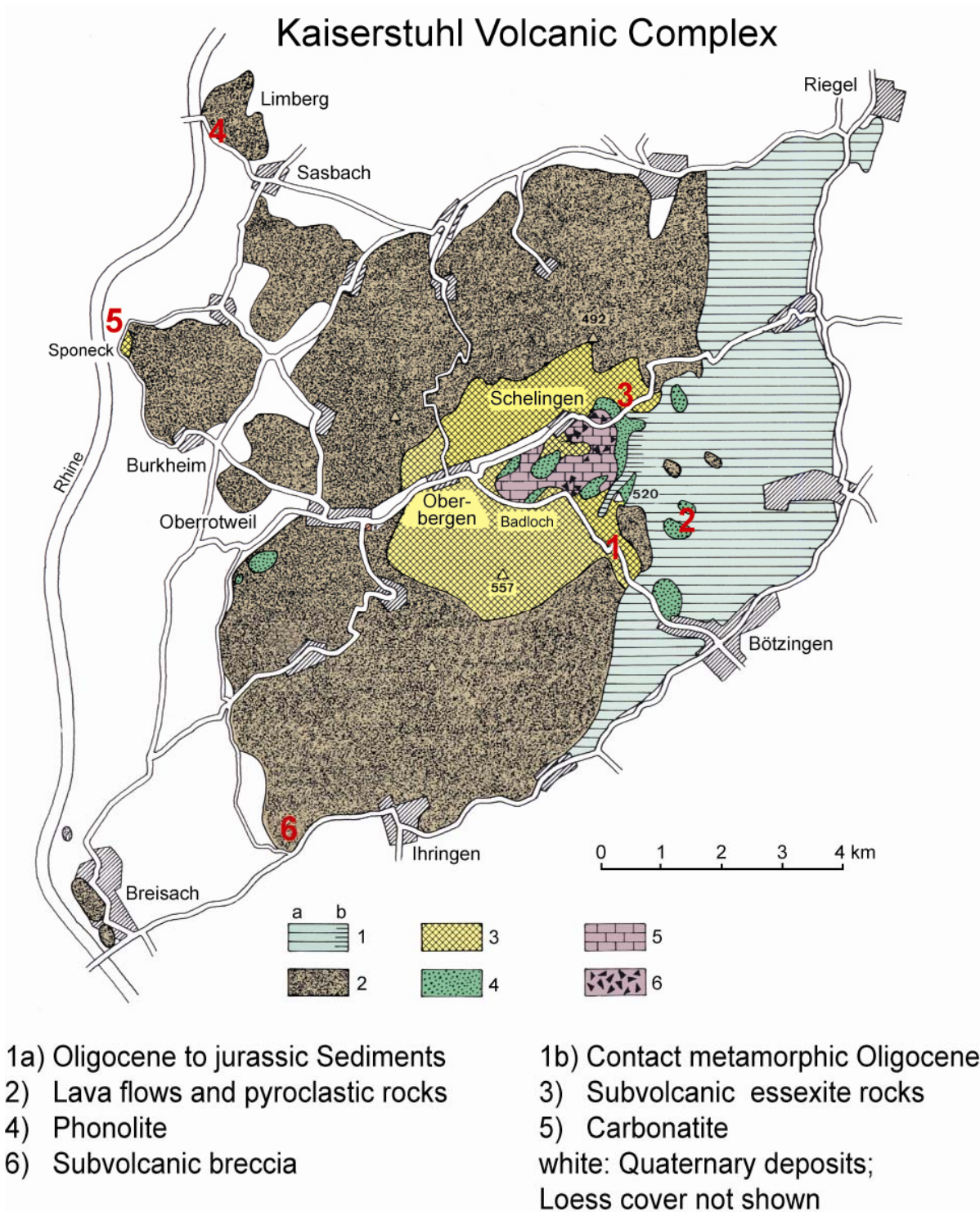


Fig. 5: Simplified geological map of the Kaiserstuhl volcanic complex.
Red numbers 1-6 refer to stops during the field trip

Tephrites and their subvolcanic equivalents form by far the largest volume of igneous rocks. Carbonatites are generally grouped with the phonolitic family, because of similarities in their trace element patterns. There is a general age sequence with more evolved intrusive rock types at the final stages. Carbonatites appear late in the evolution. Associated with carbonatites is the specific dike rock called "bergalite". This melilitic rock type does not enter readily one of the above rock families.

Three types of carbonatites can be distinguished based on the geological setting and petrography:

1. Middle to coarse-grained calcite-carbonatite (soviet) form subvolcanic intrusions in the center. Their outcrop area is about 1 km².
2. Numerous dikes are alvikites, generally less than 1 m, often only one or a few centimeters thick. Comb-layer crystallization is a general feature and has been interpreted to indicate rapid cooling (Katz & Keller 1981). Some of these dikes have porphyritic textures.
3. Extrusive carbonatites are similar to the alvikitic dikes. The droplet-shaped lapilli, "Pele's tears", of the carbonatitic melt are a rather unique occurrence of volcanic calcite carbonatites.

Calcite, magnetite, apatite are the essential mineral in the carbonatites. Carbonates are almost exclusively calcitic, only a few late-stage veins with ankeritic dolomite have been described. Small amounts of the following accessories may be present: forsterite, diopsidic augite, melanite, Nb-perovskite, pyrochlore, baryte, pyrite and rarely zirkonolite, baddeleyite, zirkon. Several micas are known: phlogopite, and Ba-phlogopite.

High pressure phases or mantle nodules are restricted in the Kaiserstuhl to two occurrences: to olivine nephelinites from the Limberg area near Sasbach and to pipe-breccias associated with the carbonatite in the center of the volcano. Tephrites are typically devoid of mantle or high pressure phases.

The general evolution of the Kaiserstuhl volcano is summarized as follows (Keller, 1984):

- The volcanological evolution starts about 17-18 Ma with the formation of a large, but complex tephritic cone. Its lava flows and pyroclastic rocks overly tertiary sediments in the eastern part of the Kaiserstuhl. Tephrite volcanism lasted all along the time of igneous activity of the complex
- Large subvolcanic intrusions of tephritic magma into the volcanic edifice formed the coarser crystalline rocks, especially essexites and ledmorites, also phonolitic domes.
- A next phase is characterized by the formation of countless volcanic dikes in the subvolcanic center. Finally such intersecting dikes of highly differentiated compositions seem to predominate in the center.
- Carbonatite intrusions occur towards the end of the evolution in the center of the complex. They are accompanied by the formation of ultramafic breccias which fill pipe structures at the subvolcanic level. These are termed "intrusive breccias". Small alvikitic dikes are the final magmatic intrusions in the center of the complex.
- The youngest volcanic activity in the volcanic complex took place in the parasitic Limberg volcano with its olivine nephelinites, limburgites and final tephrites.

Geochemistry and Petrology

Table 1 summarizes chemical data for the major rock types (Wimmenauer, 2003). All Kaiserstuhl rocks are undersaturated and plot without exception in the alkaline field of the total alkali vs. silica diagram (Fig. 6). Differences in the chemical composition and isotopic composition, for example enrichment of incompatible elements or differentiation indices like Mg-value, indicate the following genetic grouping (Keller, 1984):

Olivine nephelinites are the only primitive and possible primary magmas exposed in the Kaiserstuhl complex itself, but olivine melilitites of similar age occur in the immediate vicinity of the Kaiserstuhl. Limburgite has a basanitic composition and is more fractionated. All these rock types are sodium-rich and have high Na₂O/K₂O ratios (>1).

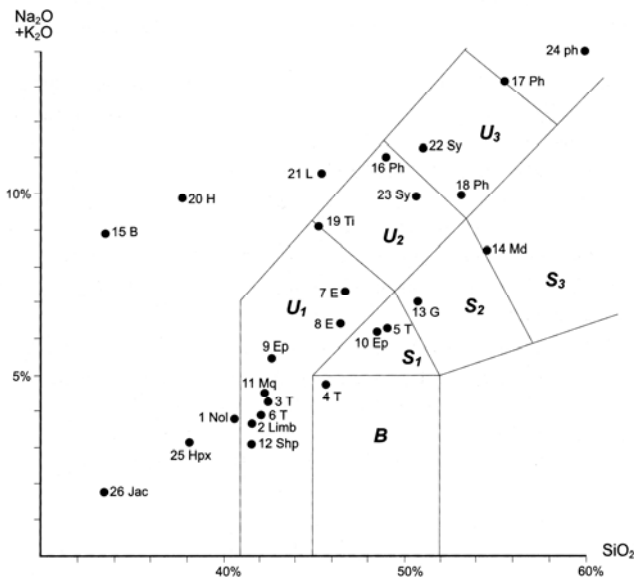


Fig. 6: Composition of the major rock types from the Kaiserstuhl volcanic complex. Total alkali-SiO₂ diagram (after Le Maitre, 1989.) The numbers refer to number of the analysis given in Table 1.

Extreme fractionation results in the formation of bergalites, which are ultrabasic, severely silica undersaturated melilites (melilite-nepheline-sodalite-perovskite-biotite-calcite-bearing). They have high Na₂O/K₂O ratios, similar to those of olivine nephelinites. These rocks have the highest concentrations of P, Nb, Sr, Ba, Y, and REE. In space and time, and by their high LIL element concentrations and by their high carbonate content, bergalites are an important link between the silicate magma series and carbonatites.

Panina et al. (2000) determined the composition of clinopyroxene and melt inclusions in this mineral from the alkaline rock series of the Kaiserstuhl. The homogenization temperatures of inclusions in the clinopyroxene systematically decrease in the sequence from olivine nephelinite (1200-1000°C) to leucite tephrite and porphyritic essexite (1170-1050°C), phonolite (1060-1040°C), and tinguaita (1140-900°C) supporting the differentiation sequence outlined above.

Strontium, Nd and Pb isotope data of the alkaline volcanic rocks and carbonatites distinguish three diverse groups which are correlated to the different magmatic series (Schleicher et al., 1990, 1991):

1. Unfractionated primary mantle melts (olivine nephelinites, olivine melilitites) and nepheline basanites: $^{87}\text{Sr}/^{86}\text{Sr} = 0.7032\text{-}0.7040$; $\epsilon_{\text{Nd}} = 4.18\text{-}5.03$; $^{206}\text{Pb}/^{204}\text{Pb} = 18.95\text{-}18.98$. Geochemically, these rocks correspond to the Na-dominated magma series.
2. Fractionated alkaline rocks (tephrites, phonolites): $^{87}\text{Sr}/^{86}\text{Sr} = 0.7039\text{-}0.7051$; $\epsilon_{\text{Nd}} = 2.07\text{-}3.96$; $^{206}\text{Pb}/^{204}\text{Pb} = 19.10\text{-}19.42$). These rocks belong to the K-dominated magma series.
3. The group comprises carbonatites, bergalites, hauynophyres ($^{87}\text{Sr}/^{86}\text{Sr} = 0.7036\text{-}0.7040$; $\epsilon_{\text{Nd}} = 2.81\text{-}4.04$; $^{206}\text{Pb}/^{204}\text{Pb} = 19.24\text{-}19.66$).

These data suggest that the diverse magma series are derived from different mantle sources. Interaction of asthenospheric melts with subcontinental lithospheric mantle during their ascent, the mixing of derivative melts with fresh magma batches, contamination of the tephrite and phonolite magma with crustal components may explain the different grades of enrichment of incompatible elements and the variations in the isotopic signatures of these melts. For the carbonatites a genetic linkage to olivine nephelinites is most likely. This implies an asthenospheric origin of the carbonatite melts.

Tephrites (including essexitic dikes and intrusions) are more fractionated (Mg#: 45-55) and have distinctively lower Na₂O/K₂O ratios of about 1. However, they are only slightly enriched in incompatible elements compared to olivine nephelinites. Therefore, it is hypothesized that tephrites cannot be derived by simple fractional crystallization from the olivine nephelinite, and have a less undersaturated, probably basanitic parent (Keller, 1984).

Phonolite and tinguaites are highly fractionated alkali rocks as indicated by their even lower Mg# and higher enrichment of incompatible elements. They could represent advanced crystallization products of the tephritic melt.

The isotopic data point to a metasomatic overprinting of the upper mantle below the Kaiserstuhl. This metasomatism is thought to be directly connected with the updomed mantle structure in this region.

From a geochemical and petrological point of view five major rock types can be distinguished in the Kaiserstuhl volcanic complex (Wimmenauer, 2003; Keller, 1984; Schleicher et al., 1990; 1991):

1. Tephrites and essexites represent the dominant volume of erupted magma. They form a potassic series, are fractionated, and their fractionation occurred in a shallow crustal level.
2. Phonolitic, tinguaitic, mondhaldeitic and other dike rocks with high K/Na show a distinct LIL element enrichment and represent further fractionation stages of the tephritic magma.
3. Olivine nephelinites and coeval olivine nephelinites in the graben close to the Kaiserstuhl represent the most primitive melts of Na-alkaline characteristics.
4. Ultrabasic, highly fractionated melilitites (bergalites) belong to the sodic clan and are considered to represent a transitional evolutionary stage towards carbonatites.
5. Carbonatites occur as sövitic intrusions, alvikitic dikes and as volcanic rocks.

Table 1: Chemical analyses of the major rock types from the Kaiserstuhl volcanic complex

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17
SiO ₂	40,79	41,66	43,55	45,67	49,12	42,10	46,80	46,64	42,89	48,57	42,43	41,70	50,94	54,63	33,63	49,05	55,79
TiO ₂	2,55	3,33	3,05	2,99	2,55	3,49	2,65	2,66	2,86	2,37	2,65	2,27	1,55	1,38	1,73	0,35	0,31
Al ₂ O ₃	10,16	11,45	15,34	15,13	15,51	12,89	15,65	15,15	14,52	14,63	13,83	14,37	17,27	18,78	13,30	17,99	19,45
Fe ₂ O ₃	5,45	5,16	4,50	4,33	3,88	6,72	5,28	5,56	5,88	4,44	2,08	6,35	5,69	4,40	5,69	2,70	2,77
FeO	5,60	6,72	6,85	7,21	6,47	5,29	5,42	4,13	5,72	5,43	6,48	4,76	2,01	2,29	4,43	1,30	0,67
MnO	0,16	0,15	0,18	0,18	0,17	0,17	0,18	0,18	0,18	0,16	0,17	0,24	0,14	0,14	0,43	0,28	0,23
MgO	14,50	11,55	5,37	5,98	4,72	6,93	4,55	4,46	5,91	4,91	8,03	6,11	2,98	1,79	3,78	0,71	1,63
CaO	11,50	12,23	12,21	13,00	10,66	13,39	11,00	10,15	13,07	10,80	12,61	14,75	7,42	5,48	13,79	7,78	2,13
Na ₂ O	2,99	2,50	2,16	2,85	2,97	2,67	3,87	2,69	2,93	2,84	2,09	0,84	2,16	3,49	6,93	6,22	7,94
K ₂ O	0,83	1,20	2,15	2,05	3,34	1,28	3,47	3,76	2,55	3,33	2,39	2,20	5,03	5,01	2,04	4,84	5,44
P ₂ O ₅	0,77	0,70	0,70	0,61	0,59	0,93	0,55	0,51	0,64	0,75	0,54	0,77	0,48	0,43	0,86	0,14	0,06
H ₂ O	2,90	2,30	3,78	-	-	3,29	1,29	2,36	2,54	1,63	2,80	3,94	3,15	1,50	4,90	5,40	1,16
CO ₂	0,85	0,75	0,07	-	-	0,09	0,12	0,92	0,21	0,04	0,18	0,87	1,08	0,07	6,71	2,52	0,22
Summe	99,05	99,70	99,91	100,00	100,00	99,24	100,83	100,01	99,90	99,90	99,28	99,17	99,90	99,12*	99,29	99,28	98,74
Cr	488	348	9	42	25	107	20	17	24	34	256	65	14	7	8	-	-
Ni	386	183	16	23	19	51	25	25	18	23	59	30	9	2	2	-	-
V	242	316	352	369	331	332	314	322	349	323	370	340	240	240	430	88	41
Zr	193	222	221	231	227	315	227	290	235	199	180	300	275	294	420	680	680
Nb	65	67	97	91	89	99	102	124	96	85	72	120	150	94	548	595	595
Sr	859	796	807	906	917	1115	924	1368	1062	1123	1117	2070	1340	1066	3003	2003	2003
Ba	753	494	651	821	972	988	789	793	7757	1172	792	1300	1530	1623	2320	1889	1889
Rb	20	18	104	111	127	121	110	119	80	128	107	75	200	158	63	225	225

	18	19	20	21	22	23	24	25	26	27	28	29	30	31	32	33	34	35
SiO ₂	53,12	45,38	37,75	45,55	51,10	50,80	60,09	38,30	33,80	71,30	61,90	24,00	1,37	3,98	2,27	1,09	3,43	1,27
TiO ₂	0,20	0,42	1,37	1,34	0,37	0,19	0,33	3,80	3,55	0,90	0,09	0,56	0,05	0,07	0,09	0,00	0,31	0,16
Al ₂ O ₃	19,62	18,45	16,52	14,10	18,27	18,17	18,58	11,88	7,24	14,75	14,60	6,37	0,25	1,50	0,88	0,17	0,35	0,54
Fe ₂ O ₃	3,14	4,74	6,03	6,95	1,42	1,97	3,64	13,90	8,77	0,53	0,83	2,68	3,18	1,99	4,59	0,91	2,97	2,24
FeO	0,74	1,40	2,61	2,12	0,96	0,43	0,10	7,61	7,61	3,95	0,16	0,93	-	-	-	-	-	-
MnO	0,25	0,34	0,31	0,22	0,13	0,15	0,09	0,16	0,47	0,14	0,11	0,86	0,42	0,25	0,45	0,46	5,04	0,32
MgO	0,41	2,29	2,45	1,63	0,63	0,28	0,18	9,03	10,29	1,57	0,38	2,04	3,05	2,28	1,28	0,31	9,43	0,58
CaO	5,08	8,21	10,63	11,84	6,96	9,99	0,57	11,75	20,15	1,36	7,82	34,85	48,60	49,50	47,90	52,20	17,73	52,36
Na ₂ O	3,63	5,27	7,22	5,50	1,35	5,18	1,37	1,08	1,08	3,07	2,78	1,52	0,10	0,29	0,03	0,00	0,21	0,12
K ₂ O	6,37	3,89	2,73	4,98	10,02	4,80	12,63	2,08	0,65	2,53	3,40	1,42	0,03	0,63	0,05	0,00	0,60	0,05
P ₂ O ₅	0,07	0,09	0,55	0,57	0,05	0,06	0,08	0,75	1,61	0,09	0,98	2,81	2,87	3,52	1,98	1,23	0,08	1,47
H ₂ O	5,09	4,87	4,43	1,26	1,85	3,03	3,10*	4,37*	4,37*	0,49	1,12	1,39	-	0,57	-	-	0,14	1,85
CO ₂	2,52	3,26	3,56	2,34	6,25	3,82	-	-	5	5,53	21,10	37,20	34,77	37,30*	41,10*	28,55	37,40	
Summe	99,22	98,89	98,93	99,65	99,36	99,18	100,22	99,57	99,59	100,68	99,70	100,54	97,12	99,90	96,82	97,47	80,09	98,36
Cr	-	-	12	-	-	-	(20)	128	275	-	-	-	-	-	-	-	-	-
Ni	< 10	(44)	11	(46)	-	-	(< 5)	< 10	13	-	-	-	-	-	-	-	-	-
V	46	(390)	298	(546)	240	-	(195)	-	-	-	-	-	90	-	160	130	331	55
Zr	568	-	463	-	-	320	(190)	-	-	385	65	233	180	-	260	40	0	49
Nb	595	(370)	307	(481)	920	130	(530)	16	18	34	53	502	2400	(8300)	1500	890	67	565
Sr	6069	(1760)	2818	(2165)	2750	3370	(2900)	610	310	300	100	5205	9000	-	7660	7990	0,95 %	4160
Ba	2088	(1250)	2719	(2405)	3020	2250	(7000)	850	1800	405	70	1090	650	-	800	960	19,30 %	1290
Rb	351	(274)	345	(266)	300	170	(180)	14	14	358	514	96	-	-	-	-	-	-

Type and source of samples analyzed in Table 1 (Wimmenauer, 2003):

- 1: Olivine nephelinite, Lützelberg
- 2: Limburgite L₁, Limberg Quarry I
- 3: Tephrite lava bombe, Blankenhornsberg
- 4: Average of 80 tephrites group (main type)
- 5: Average of 20 tephrites group IV (fractionated tephrites)
- 6: Gestein des Lavastromes (phonolithischer Tephrit), Limberg Stör. VII.
- 7: Essexite, Eichholz near Oberbergen
- 8: Essexite, Krummer Graben near Oberbergen
- 9: Essexite dike, Oberbergen
- 10: Essexite dike, Föhrenberg near Ihringen
- 11: Olivine monchiquite, Oberrotweil
- 12: Shonkinite, Horberig
- 13: Gauteite, Horberig
- 14: Mondhaldeite, Föhrenberg near Ihringen
- 16: Bergalite, Hessleterbuck near Oberbergen
- 16: Phonolite, Fohberg near Bötzingen-Oberschaffhausen
- 17: Phonolite, Kichberg near Niederrotweil
- 18: Phonolite dike, Horberig near Oberbergen
- 19: Tinguait, Strümpfekopf near Bickensohl;
- 20: Hauynophyre, Drill core Steinreiß, 16,5 m;
- 21: Ledmorite, Badberg-Südhang;
- 22: Foidal syenite, Research drill core Kaiserstuhl
- 23: Foidal syenite, xenolith in phonolite, Kirchberg near Niederrotweil
- 24: "Trachyte", Ohrberg near Schelingen
- 25: Pyroxene-Hornblendite ("Yamaskit") from polygenetic pyroclastic rock, Nonnensohl near Oberschaffhausen
- 26: Pyroxenite ("Jacupirangit") from intrusive Breccie, Lehr near Schelingen
- 27: Granite, from polygenetic pyroclastite rock, Eichstetter Vulkan
- 28: Syenitic fenite, from polygetic pyroclastic rock, Eichstetter Vulkan
- 29: Ringitic fenite, Tuff t3, Limberg Quarry VI
- 30: Sövite, Orberg near Schelingen, Quarry V;
- 31: Sövite, research drill core Kaiserstuhl
- 32: Alvikite with comb structure, Kirchbuck near Oberbergen, marginal zone
- 33: Alvikite with comb structure, Kirchbuck near Oberbergen, central zone;
- 34: Magnesio-carbonatite with baryte, drill core KB2, Schelingen
- 35: Carbonatitic lapilli, Henkenberg near Niederrotweil

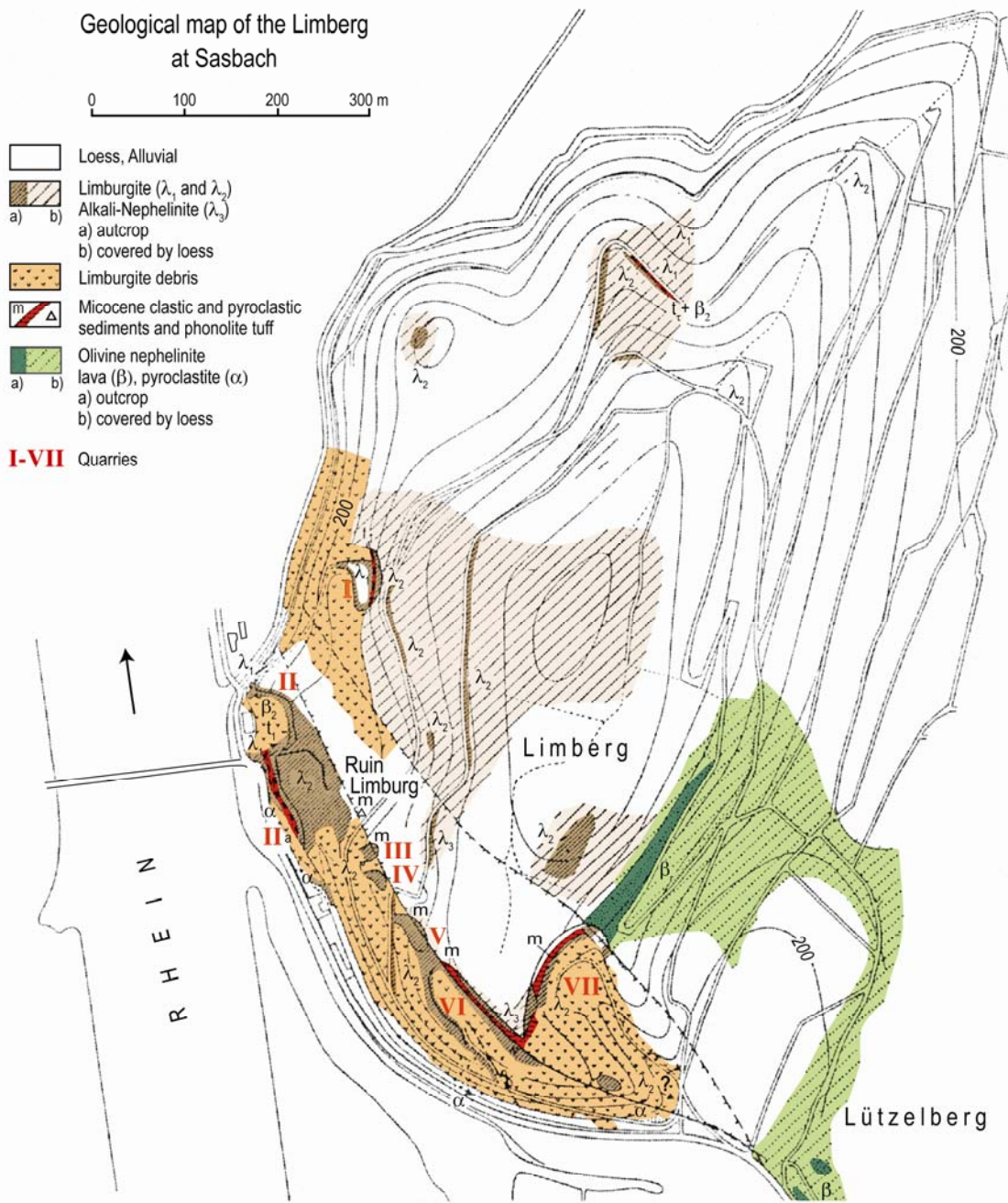


Fig. 7: Detailed geological map of the Limberg-Lützelberg volcanic complex (after Wimmenauer, 2003)

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B. Geology and Petrology of the Schwarzwald

Modified after: Angelika KALT, Rainer ALTHERR, Michael HANEL (2000) The Variscan basement of the Schwarzwald. *Beih. z. Eur. J. Mineral.* Vol. 12,2000 No.2, S. 1 - 43

1. OVERVIEW: THE MAIN TECTONIC UNITS OF THE SCHWARZWALD

The crystalline basement of the Schwarzwald consists of four tectonometamorphic complexes. These are from north to south (Fig. 2) the Baden-Baden Zone (BBZ), the Central Schwarzwald Gneiss Complex (CSGC), the Badenweiler-Lenzkirch Zone (BLZ) and the Southern Schwarzwald Gneiss Complex (SSGC). All four units are composite and were intruded by Variscan granitoids (Fig. 1,2). The BBZ and the BLZ are fault zones with tectonic slivers of non- to high-grade metamorphic rocks that dip steeply under the allochthonous high-grade complex of the CSGC from the north (BBZ) and from the south (BLZ). The high-grade SSGC is separated from the BLZ by a major normal fault, juxtaposing high-grade metamorphic rocks in the south with nonmetamorphic sediments and volcanic rocks to the north (Fig. 2). Towards the east, the basement of the Schwarzwald is overlain by Permian and Triassic sediments. The western margin of the Schwarzwald is formed by the Tertiary Rhinegraben.

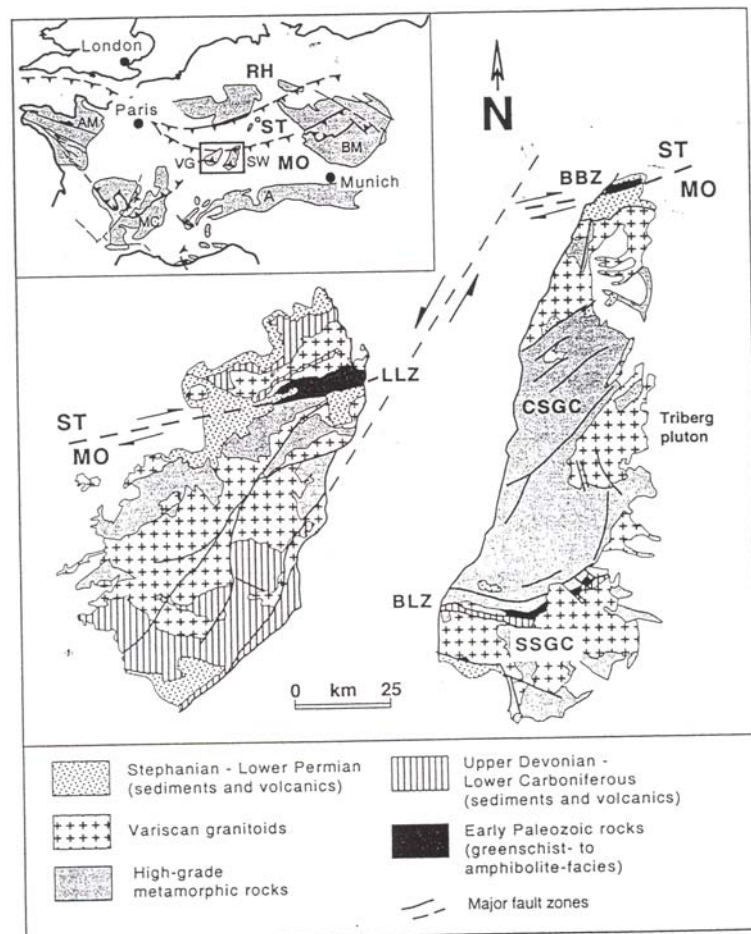


Fig. 2. Geological sketch map of the Schwarzwald and the Vosges, modified after Echtler & Chauvet (1991/92). The inset shows the location of Schwarzwald and Vosges within the Variscan belt in Europe. Abbreviations are A: Alps, AM: Armorican Massif, BBZ: Baden-Baden Zone, BLZ: Badenweiler-Lenzkirch Zone, BM: Bohemian Massif, CSGC: Central Schwarzwald Gneiss Complex, LLZ: Lalaye-Lubine Zone, MC: Massif Central, MO: Moldanubian Zone, RH: Rhenohercynian Zone, SSGC: Southern Schwarzwald Gneiss Complex, ST: Saxothuringian Zone, SW: Schwarzwald, VG: Vosges.

Classically, CSGC, BLZ and SSGC are assigned to the Moldanubian zone of Kossmat (1927), whereas the BBZ is seen as part of Kossmat's Saxothuringian zone (e.g. Franke, 1989; Fig.3). Although separated only 30 km by the Tertiary Rhinegraben, the metamorphic and magmatic basement outcrops of Schwarzwald and Vosges are not easily correlated. Displaced 30 km to the north by a sinistral shear zone, the BBZ continues to the west as the Lalaye-Lubine Zone (Fig. 1). A simple correlation of the CSGC with the Central Vosges is not possible as both consist of various metamorphic complexes. Devonian to Carboniferous low-grade to non-metamorphic sediments and volcanic rocks that form slivers within the narrow BLZ of the Schwarzwald cover vast areas in the Southern Vosges (Fig. 2). No equivalents of the SSGC are found in the Vosges. Towards the east, metamorphic rocks similar to some of the CSGC are found as xenoliths within the Tertiary volcanic field of Urach and within the Ries crater. Also, a few lithological correlations between parts of the CSGC and the south-western Bohemian Massif are possible. Recent paleomagnetic studies of Edel (2004) suggest that the whole Moldanubian zone, from the Massif Central in the west to the Sudetes in the east, moved as one block since the early Carboniferous.

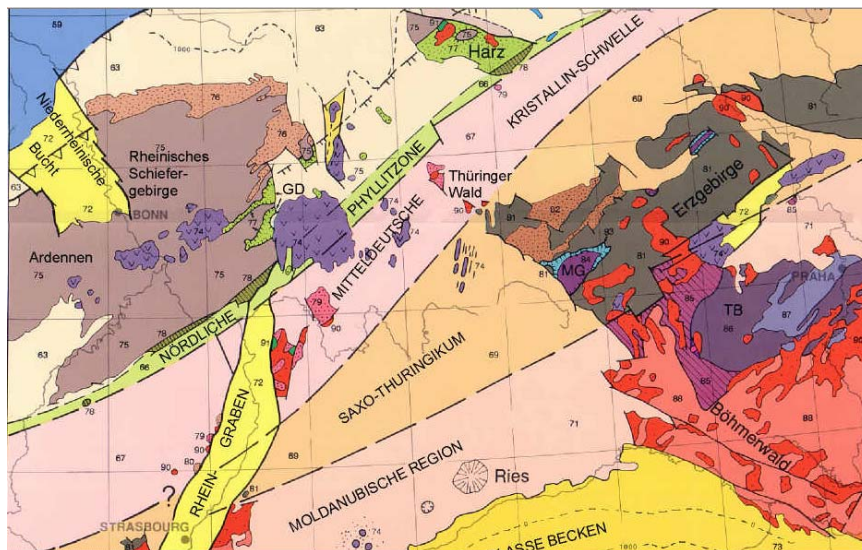


Fig. 3 Terranes of the Variscan Mountain Chain

a. THE BADEN-BADEN ZONE

The BBZ is a narrow SW-NE trending structure at the northernmost tip of the Schwarzwald basement (Fig. 2) that is generally assigned to the Saxothuringian zone (e.g. Franke 1989). Most of the BBZ is covered by undeformed sediments and volcanic rocks of Stephanian to Lower Permian age (e.g. Sittig, 1974, 1983; Fröhler & Lebede, 1994; Hess et al., 1983; Fig. 2). Within the Permo-Carboniferous cover, isolated outcrops of older metamorphic and magmatic rocks occur (Fig. 2). The metamorphic rocks were divided into three groups by Wickert et al. (1990) Group A consists of gneisses and amphibolites that equilibrated at high-temperature low pressure (HT-LP) conditions that are similar to those of the final metamorphic overprint in the CSGC (730-780°C at 0.40-0.45 GPa; Kalt, 1995; see section on the CSGC). Group B comprises garnet- and kyanite-bearing schists and quartzites that experienced metamorphic conditions of 0.9 GPa and 630-670°C (Wickert et al., 1990; Fig. 3). Group C consists of metasediments and greenschists, metamorphosed at approximately 0.2 GP and 450°C, and slices of deformed two-mica granite form unit D (Wickert et al., 1990; Fig. 3).

During Variscan convergence, group C rocks were accreted along NE-SW trending, SE dipping mylonitic shear planes (Wickert et al., 1990; Fig. 2). Group A and B rocks do not show this mylonitic foliation. They were imbricated along roughly E-W trending cataclastic strike-slip faults. 40 Ar/39 Ar ages of 334.5 ± 1.6 Ma for muscovite and 331.4 ± 2.0 Ma for biotite from group B schists (Altherr, unpubl. data) give maximum estimates for the age of this cataclastic accretion. The latter was followed by the intrusion of a biotite-hornblende granite (Baden-Baden granite) in the late Early Carboniferous (see below) that caused contact metamorphism in group C rocks. Late, roughly E- W trending and steeply S-dipping shear zones, also in the granite, suggest exhumation in the Late Carboniferous

within an extensional regime (Wickert et al., 1990).

The Baden-Baden granite belongs to a suite of Early Carboniferous high-K, calc-alkaline I-type plutonic rocks that is mainly found in the northern Vosges (Altherr et al., 2000). Within this suite, an older group of diorites, granodiorites and granites is distinguished from a younger group of more evolved granites including the Baden-Baden granite. K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ hornblende and biotite ages for the I-type plutons of the northern Vosges range from 330 ± 4 to 325 ± 4 Ma (Boutin et al., 1995; Altherr et al., 2000). The metaluminous to slightly peraluminous plutons of the I-type suite are cut by peraluminous two-mica granites. The emplacement age of the Kagenfels granite in the Vosges, forming the largest of these S-type granitic bodies, is constrained by single zircon Pb evaporation ages to 331 ± 5 Ma (Hess et al., 1995). Geochemical and isotopic evidence suggests that the diorites were most likely derived from enriched mantle sources, whereas the granodiorites and granites probably formed by dehydration melting of heterogeneous metasedimentary sources at increasing depths (Altherr et al., 2000).

b. THE CENTRAL SCHWARZWALD GNEISS COMPLEX

Detailed structural and petrological studies have shown that three tectonometamorphic units can be distinguished in the CSGC (Wimmenauer, 1980, 1984; Hanel & Wimmenauer, 1990; Kalt et al., 1999a) that are often separated by mylonitic shear zones (Fig. 4). Common to all three units is the lithological dominance of metapsammitic gneisses and a HT-LP metamorphic overprint at conditions of approximately $730\text{-}780^\circ\text{C}$ / $0.40\text{-}0.45$ GPa (Kalt et al., 1994a; Kalt, 1995) at approximately 330 Ma (Kalt et al., 1994a; Lippolt et al., 1994). The units differ in their lithologies, in their metamorphic evolution before 330 Ma (Fig. 5) and most likely also in their premetamorphic history (Fig. 6).

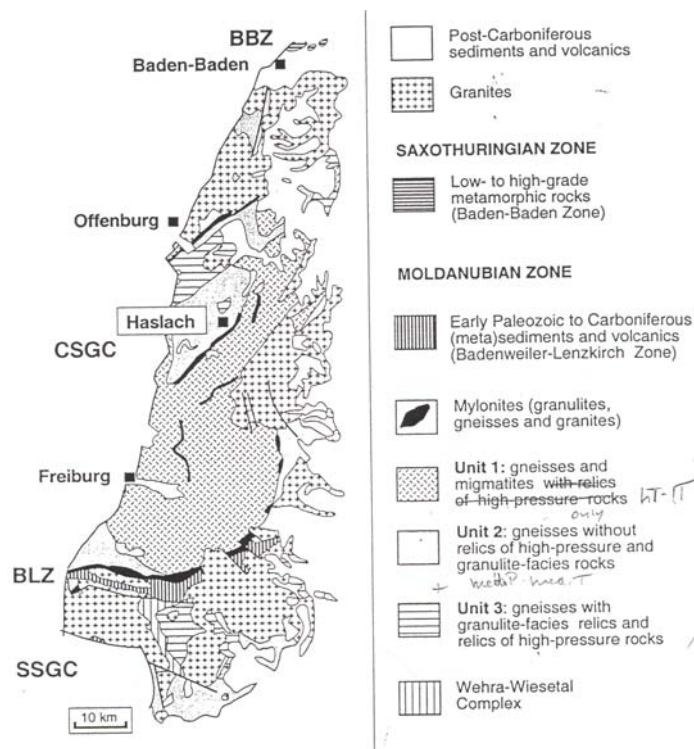


Fig. 4. Simplified geological map of the Schwarzwald, showing the distribution of the three tectonometamorphic units, mylonite zones and the Wehra-Wiesetal complex. Abbreviations are BBZ: Baden-Baden Zone, BLZ: Badenweiler-Lenzkirch Zone, CSGC: Central Schwarzwald Gneiss Complex, SSGC: Southern Schwarzwald Gneiss Complex.

Monotonous unit with high-pressure relics (unit 1)

The main rock types in unit 1 are gneisses and migmatites. Their petrography has been described by various authors (e.g. Mehnert, 1953, 1957, 1962, 1963; Wimmenauer, 1984; Wimmenauer et al., 1989). The phase assemblage of these rocks is biotite + plagioclase + quartz ± alkali feldspar ± garnet ± sillimanite ± cordierite. Metamorphic conditions are approximately 730-780°C / 0.40-0.45 GPa (Kalt et al., 1994a; Fig. 5). In the widespread psammitic lithologies, there is textural evidence for the formation and crystallisation of partial melts during HT-LP metamorphism (Mehnert, 1953, 1957, 1962, 1963), although limited compared to the more rare pelitic lithologies (Mazurek, 1992; Kalt et al., 1994a).

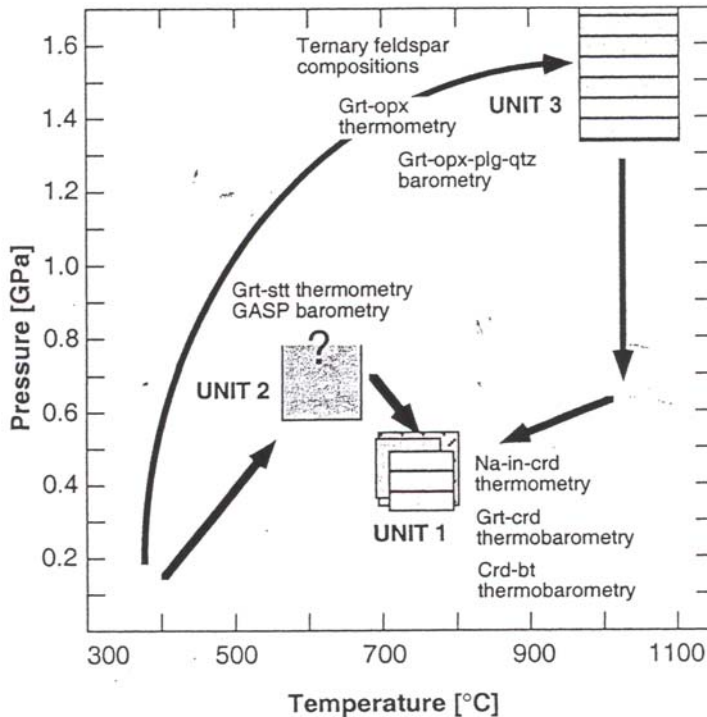


Fig. 5. P-T diagram showing the pressure temperature evolution of tectonometamorphic units 1-3 in the Schwarzwald. Unit 1 is a monotonous high-temperature low-pressure unit with no indications of an earlier metamorphic stage that contains numerous eclogite and peridotite lenses. Unit 2 is a monotonous high-temperature low-pressure unit with relics of a medium-temperature medium-pressure stage. Unit 3 is a lower crustal unit with a high-temperature low-pressure overprint. The diagram is based on Rehfeld (1983), Rohr (1990), Kalt et al. (1995), Kalt et al. (1999b) and Marschall et al. (2000).

Concordant, conventionally obtained U-Pb ages on monazite separates reflect the peak of HT-LP metamorphism. They scatter between 329 ± 1 and 332 ± 1 in the CSGC (235U/207Pb ages) and coincide with the ages obtained by Rb-Sr thin slab dating (Kalt et al., 1994a). Growth of zircons and zircon rims was obviously restricted to rocks where melts or fluids were present (Kalt et al., 2000). $^{40}\text{Ar}/^{39}\text{Ar}$ plateau ages of hornblende from amphibolites that record cooling below approximately 500°C scatter from 324 ± 2 Ma to 347 ± 3 Ma (Lippolt et al., 1994).

The results of palynological studies and of U-Pb dating of zircon from paragneisses by SHRIMP (Sensitive High Resolution Ion Microprobe) allow for some constraints to be placed on sedimentation ages. Acritarchs in one gneiss from the type locality for kinzigite (see above) suggest a Neoproterozoic age of sedimentation (1000-540 Ma B.P.; Montenari, 1996), while chitinozoans and acritarchs in other paragneisses point to Late Paleozoic sedimentation (Hanel et al., 1999a, b). The U-Pb ages of single domains in complex zircons record various events of zircon growth (Kober et al., 2000a). The oldest

inherited cores are of Achaean (2.5-2.9 Ga) and Early Proterozoic (1.8-2.0 Ga) age (Fig. 6). Most of the studied zircons grew during magmatic events around 540-650 Ma B.P. (Fig. 6) and experienced no later overgrowth or resorption. In paragneisses, there is only limited evidence for zircon growth during Variscan HT-LP metamorphism.

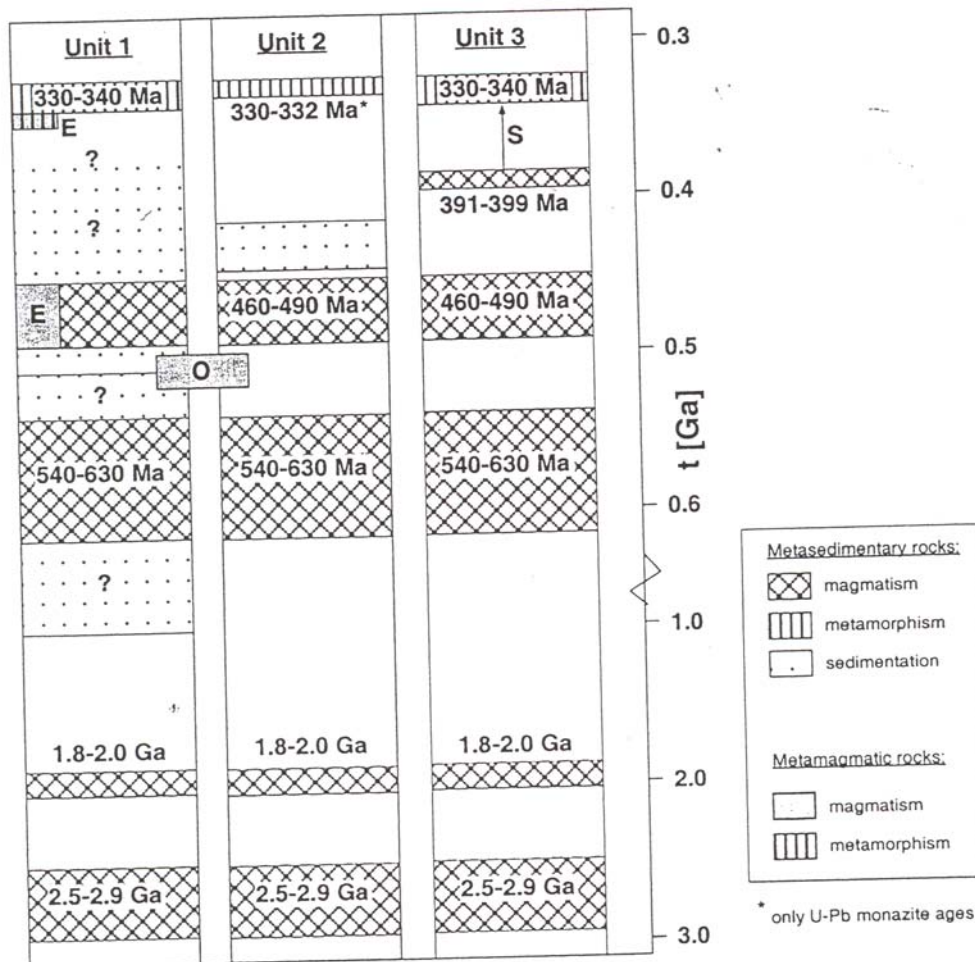


Fig. 6. Compilation of SHRIMP zircon ages and palynological data (acritarchs and chitinozoans) for metasedimentary rocks (paragneisses) of the Central Schwarzwald Gneiss Complex. Data sources are Montenari (1996), Hanel et al. (1999) and Kober et al. (2000a). Additionally, magmatic ages of orthogneiss precursors (O; U-Pb, conventional zircon dating: Todt & Büsch, 1981; Pb evaporation dating: Chen et al., 2000) and of zircons contained in an eclogite as well as the age of eclogite-facies metamorphism are shown (E; Kalt et al., 1994b; Kalt et al., 2000). Sedimentation in unit 2 is quite well constrained by palynological data. For unit 1, palynological evidence from different rocks suggests sedimentation in the Neoproterozoic as well as in the Paleozoic. No palynomorphs were found in unit 3, but sedimentation (S) must be younger than the youngest magmatic zircons contained in the metasedimentary rocks.

The high-pressure relics in unit 1 are isolated, strongly retrogressed lenses of eclogite, spinel peridotite, garnet-spinel peridotite and garnet websterite. The primary mineral assemblage in the eclogites is garnet + omphacite + rutile ± quartz ± kyanite ± clinozoisite ± phengite (Klein & Wimmenauer 1984; Stenger et al., 1989). The results of thermometry range from 600 to 750 °C, and minimum pressure estimates are between 1.2 and 1.6 GPa (Klein & Wimmenauer 1984; Kalt et al., 1994b). The age of high-pressure (HP) metamorphism has been precisely determined by Pb evaporation dating on zircons (Kalt et al., 1997) and U-Pb SHRIMP dating on zircons (343 ± 2 Ma; Kalt et al., 2000). The eclogites display various decompression, cooling and hydration reactions under granulite, amphibolite and even greenschist facies conditions (Klein & Wimmenauer, 1984). Most of

them appear as amphibolites in the field. The HT-LP overprint is recorded by fluid-induced zircon growth in eclogites at 331 ± 1 Ma (Kalt et al., 2000). Subsequent formation of retrograde greenschist-facies phases is documented by a Rb-Sr isochron of 325 ± 6 Ma (Kalt et al., 1994b).

The strongly serpentinised peridotites have the primary assemblage olivine + clinopyroxene + orthopyroxene \pm spinel \pm garnet \pm amphibole. Most of the samples record only one equilibration stage in the spinel or in the garnet-spinel peridotite stability field. The equilibration conditions (670 - 770 °C / 1.4 - 1.6 GPa; Kalt et al., 1995; Kalt & Altherr, 1996) are similar to those of the eclogites.

Variegated unit without high-pressure relics (unit 2)

In the CSGC, rocks of unit 2 occur mainly within a tectonic window beneath unit I northwest of Haslach and at the southern margin with the BLZ (Fig. 4). Apart from biotite-plagioclase gneisses, unit 2 consists of graphite-bearing gneisses, quartzites, marbles, calcsilicate rocks, leucocratic gneisses and amphibolites (Wimmenauer, 1980, 1984; Wimmenauer & Lim, 1988). Intimate interlayering of these different rock types may occur and evidence for partial melting is not very widespread in unit 2. Very rare staurolite-, kyanite- and garnet-bearing rocks allow for the estimation of P-T conditions. New P-T calculations suggest temperatures of 690 - 720 °C and pressures between 0.7 and 1.5 GPa (Kalt; unpubl. data; Fig. 5). Equilibration of the staurolite-garnet-kyanite gneisses at this MT-MP stage was followed by partial re-equilibration at HT-LP conditions as suggested by late cordierite and sillimanite (Rehfeld, 1983). In contrast to unit I, the amphibolites of unit 2 show no relics of a former eclogite stage. Their mineral assemblage is essentially amphibole + plagioclase + titanite \pm clinopyroxene \pm quartz. A concordant $^{235}\text{U}/^{207}\text{Pb}$ age of 333 ± 2 Ma (Kalt et al., 1994a), obtained from monazites of a leucocratic gneiss without MT-MP relics most likely dates the HT-LP overprint in unit 2. As in unit I, zircons from paragneisses hardly record this event.

The results of palynological studies and of U-Pb dating of zircon from paragneisses by SHRIMP place some constraints on sedimentation ages. Acritarchs and chitinozoans point to Ordovician / Silurian sedimentation (455 - 415 Ma B.P; Hanel et al., 1999a, b). The U-Pb ages of single domains in complex zircons record various events of zircon growth (Kober et al., 2000a). The Archean (2.5 - 2.9 Ga), Early Proterozoic (1.8 - 2.0 Ga) and Late Proterozoic to Cambrian (540 - 650 Ma) ages recording magmatic events are approximately the same as those found in paragneisses of unit 1 (Fig. 6). In addition, the paragneiss zircons of unit 2 record magmatic events around 460 - 490 Ma (Fig. 6).

In the area around Haslach (Fig. 4), particularly in the Kinzig valley, larger bodies of orthogneiss with tonalitic, trondhjemitic and granitic compositions crop out (Wimmenauer, 1984; Klein & Stenger, 1987). Some of these orthogneisses are in close spatial relation to paragneisses of unit 1, others to unit 2 and can thus not be unequivocally assigned to either unit. The intrusion of the orthogneiss precursors was dated at 520 ± 5 Ma (Todt & Büsch, 1981; Fig. 6). Slightly younger ages of 497 - 512 Ma were later obtained by conventional U-Pb and Pb evaporation dating (Chen et al., 2000).

Retrogressed granulites (unit 3)

Gneisses with relics of a granulite-facies metamorphic stage can be found in a coherent area northwest of Haslach (Rohr, 1990; Fig. 4) and in several discrete mylonitic shear zones. The shear zones either separate unit 1 from unit 3 or occur within unit 1 (Fig. 4). The gneisses are mainly of metapsammitic and metapelitic composition, more rarely metabasites and spinel peridotites occur. The metapsammitic gneisses are mainly quartzo-feldspathic with relics of ternary feldspars (now exsolved) and garnet. The metapelitic gneisses bear the granulite-facies assemblage garnet + kyanite + rutile + ternary feldspar + quartz \pm biotite. Recent investigations on these rocks have revealed a prograde evolution in the kyanite stability field, peak conditions of 950 - 1000 °C at minimum pressures of 1.5 GPa, isothermal decompression of 1.0 GPa and isobaric cooling at very low pressures in the cordierite stability field (Marschall et al., 2000; Fig. 5). Partial melts were probably present during most of the prograde and retrograde path (Marschall et al., 2000). The metabasic rocks bear evidence of a very early eclogite-facies stage (garnet + kyanite + omphacite + rutile + quartz), followed by equilibration

in the granulite facies (garnet + orthopyroxene + clinopyroxene + plagioclase + quartz) and partial re-equilibration and hydration under amphibolite facies conditions (Hanel et al., 1993). The widespread occurrence of the granulite relics suggests that unit 3 represents a lower crustal level.

Pb evaporation dating on zircons from the basic granulites yielded an age of 341 ± 19 Ma (Hanel et al., 1993). The age of the granulite-facies equilibration of metapelitic granulites was dated by SHRIMP analyses on outer zircon domains and newly formed or recrystallised zircon grains with 335-340 Ma (Kober et al., 2000a). Chen et al. (1998a, b) report a Sm-Nd age of 402 ± 3 Ma on garnet from a basic granulite and a Pb-Pb garnet-whole rock age of 380 ± 16 Ma from a 'granulitic felsite'. The ages are interpreted by the authors as 'HT and HP events at ca. 380 to 400 Ma'. The upper intercept of a U-Pb zircon discordia of 421 ± 25 Ma is interpreted as a magmatic event (Chen et al., 1998a, b). SHRIMP ages of 391-399 Ma, however, obtained on magmatic zircon domains from metapelitic granulites (Kober et al., 2000a) indicate a maximum sedimentation age in the Lower Devonian (391-399 Ma) for the sedimentary precursors and hence do not argue in favour of a metamorphic event at 380-400 Ma. As in units 1 and 2, Achaean and Early Proterozoic zircon cores are present (Fig. 6). In addition, magmatic zircon formation at 460-490 Ma is evident from SHRIMP data (Fig. 6; Kober et al., 2000a).

Granites

The metamorphic rocks of the CSGC are intruded by numerous granitoids of variable size. In the north of the CSGC, the plutons of Bühlertal, Forbach, Raumünzsch, Sprollenhaus, Wildbad, Seebach and Oberkirch form the so-called Northern Schwarzwald Granite Massif (Fig. 7). Except for the Oberkirch and Wildbad plutons, all intrusions are fairly homogeneous two-mica or only muscovite-bearing S-type granites that partly bear cordierite and/or andalusite. Their chemical and isotopic compositions are consistent with derivation from (metapelitic) crustal sources by partial melting (Kober et al., 2000b; Altherr, unpubl. data). From field relations, these two-mica granites are the youngest among the granitoids in the CSGC. Their intrusion was dated by Pb evaporation on single zircon grains at 325 ± 7 Ma (Hess et al., 2000). From field relations, the biotite granites (Wildbad and Oberkirch) are older than the two-mica granites. Their zircon and mica ages, however, are exactly the same as those of the latter (Hess et al., 2000). Thus the Northern Schwarzwald Massif as a whole shows significantly younger cooling ages than the BBZ to the north and the central CSGC to the south. As the Northern Schwarzwald Granite Massif is bordered by fault zones to the north and south (Fig. 7) it may be considered as a separate tectonic unit that can also be found to the west in the Vosges (Hess et al., 2000).

The large Oberkirch pluton is compositionally heterogeneous. At the exposure level, it consists of three different units of peraluminous, partly cordierite-bearing biotite granite, with synplutonic dioritic dikes and dioritic to tonalitic microgranular enclaves of variable size occurring in each of them (Otto, 1971/72; 1974). These field relations suggest that at least three pulses of granitic magma mingled with dioritic and tonalitic magma (Altherr et al., 1999a). Chemical and isotopic evidence suggests that the dioritic magmas were derived by partial melting from an enriched mantle source with minor assimilation of crustal material, whereas the granitic magmas originated from crustal sources (Altherr et al., 1999a). Single zircon evaporation ages are 325 ± 5 Ma (Hess et al., 2000).

The large Triberg pluton in the east of the CSGC is also a composite intrusion (Schleicher & Fritsche, 1978; Schleicher 1984, 1994). At the exposure level, four compositionally different units of granite were distinguished by Schleicher (1994). Based on geochemical data an origin by partial melting of mid to lower crustal sources is deduced for the Triberg pluton (Schleicher, 1994). A two-mica leucogranite unit contains cassiterite-bearing greisen and beryl-bearing pegmatites (Markl, 1997; Markl & Schumacher, 1996, 1997). The intrusion age of the Triberg pluton is constrained by a single zircon evaporation age of 332 ± 6 Ma (Kober, unpubl. data).

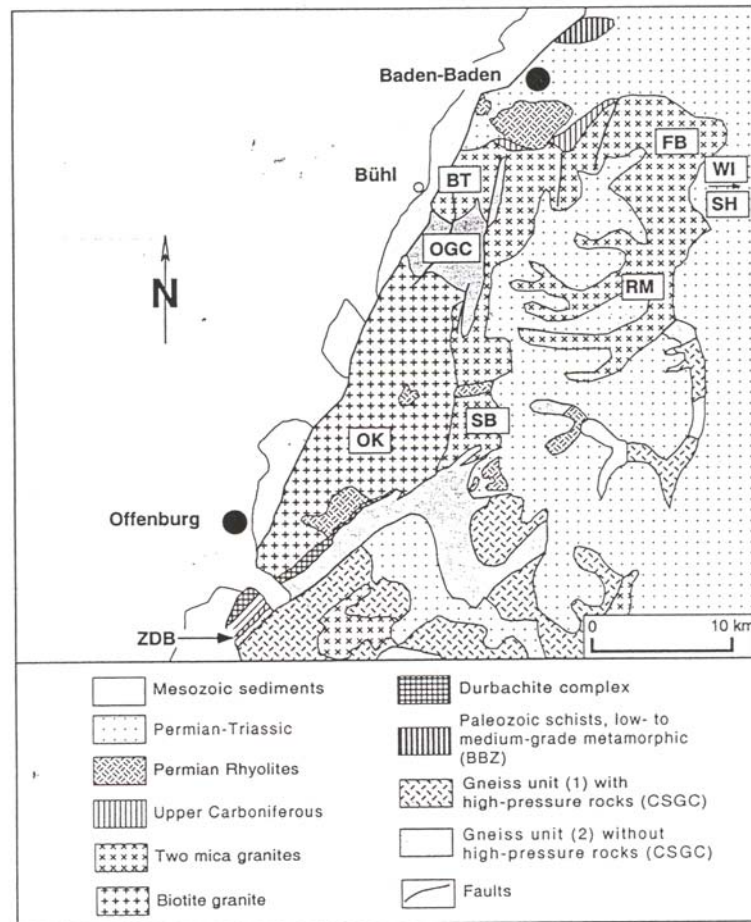


Fig. 7. Geological map of the Northern Schwarzwald Granite Massif and adjacent areas showing the locations of different granitoid plutons. Abbreviations are BT: Blühlertal granite, FE: Forbach granite, ace: Omerskopf gneiss complex, OK: Oberkirch pluton, RM: Raumünzach granite, 5B: Seebach granite, 5H: Sprollenhaus granite, WI: Wildbad granite, ZDB: Zone of Diersburg-Berghaupten.

c. THE BADENWEILER-LENZKIRCH ZONE

The BLZ is a thrust and fault zone where the CSGC has been thrust onto tectonic slivers of non- to high-grade metamorphic Paleozoic rocks in a system of SE-vergent thrusts and dextral shear zones (Altherr & Maass, 1977; Sittig, 1981; Werling & Altherr, 1987; Lüschen et al., 1987; Krohe & Eisbacher, 1988; Eisbacher et al., 1989; Wimmenauer & Schreiner, 1990; Figs. 8. The southernmost part of the CSGC comprises a heterogeneous complex of biotite and biotite-hornblende gneisses, leucocratic gneisses, amphibolites and porphyroclastic metagranites ('Randgranit'; Hoenes, 1940, 1948; Altherr & Maass, 1977; Büsch & Mehnert, 1995). To the south of the 'Randgranit' complex, the BLZ proper shows ductile deformation, with the mylonitic foliation generally dipping to the NW. Within this northern part of the BLZ, different thrust sheets can be distinguished on the basis of lithology and metamorphic grade. The metamorphic grade with respect to the premylonitic and synmylonitic state is inverse and decreases discontinuously from north to south (Altherr & Maass, 1977; Werling & Altherr, 1987; Wimmenauer & Schreiner, 1990). The southernmost parts of the CSGC ('Randgranit') and the 'Spiesshorn-Verband' (Fig. 8) were equilibrated and deformed under amphibolite-facies conditions (sillimanite + cordierite + garnet). Whereas the former belongs to unit 1, the 'Spiesshorn-Verband' can most likely be correlated with unit 2 of the CSGC (Wimmenauer & Hanel, 1997). To the south, the rocks of the 'Sengalenkopf-Verband' (Fig. 8) equilibrated at lower amphibolite-facies conditions (andalusite + cordierite) and were deformed under lower greenschist-facies conditions. The 'Sengalenkopf-Verband' frequently contains foliated metaconglomerates with

pebbles of former clastic sediments and arc-type volcanics (Emmermann & Sittig, 1975, 1977; Altherr & Maass, 1977; Werling & Altherr, 1987; Wimmenauer & Schreiner, 1990). Further to the south, the low-grade metamorphic siltstones and greywackes of the 'Verband von Bernau-Wacht' (Fig. 8) were deformed at lower greenschist-facies conditions. In the southern part of the BLZ, non-metamorphic Upper Devonian to Lower Carboniferous sediments and volcanic rocks (Sittig, 1969, 1972a, b, 1981; Burgath, 1973; Burgath & Maass, 1973; Kneidl et al., 1982; Maass et al., 1990; Glüdenpfennig & Loeschke, 1991; Glüdenpfennig, 1998) are only cataclastically deformed. In the central segment of the BLZ, slivers of deformed granite ('Münsterhaldengranit') are present between the foliated metamorphic units in the north and the non-metamorphic sediments and volcanic rocks in the south.

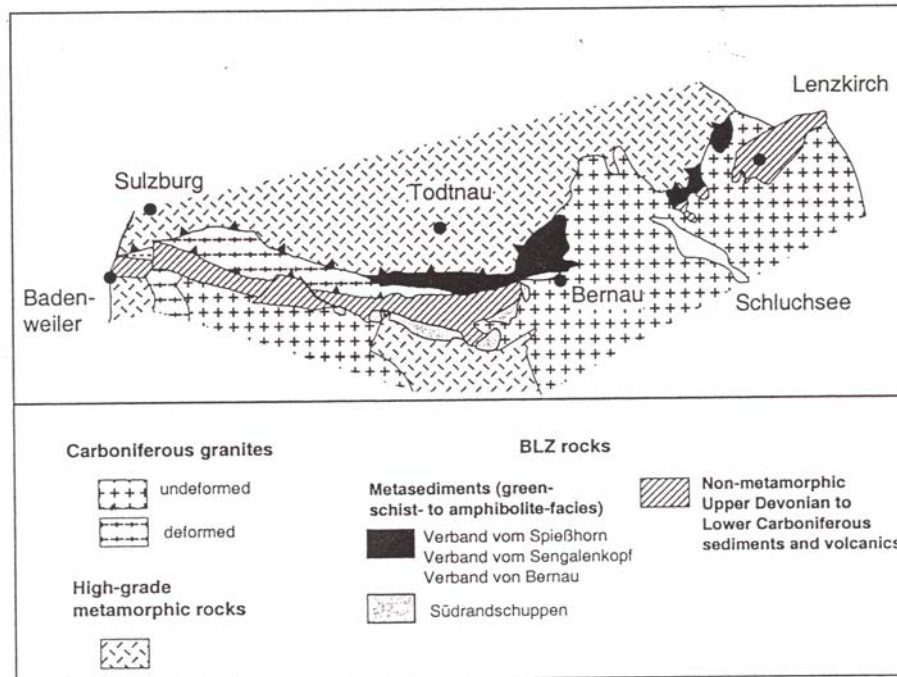


Fig. 8. Geological map of the BLZ (Werling, 1986), compiled after Metz & Rein (1958), Sittig (1969) and Burgath & Maas (1973).

Fragments of arc relics in the metamorphic rocks of the 'Sengalenkopf-Verband' and the Lower Carboniferous sediments (Vise: 350-325 Ma B.P.), the orogenic nature of the volcanic rocks (Burgath, 1973; Burgath & Maass, 1973; Altherr & Maass, 1977; Glüdenpfennig & Loeschke, 1991; Glüdenpfennig, 1998) and the possibility that Devonian fossils are reworked (Pieri-Thoma, 1987; Sittig, pers. comm.) have led to the view that the southern BLZ may represent a Lower Carboniferous syn-orogenic flysch basin, with olistostromes of arc-relics and Devonian neritic sediments (Krohe & Eisbacher, 1988; Echtler & Altherr, 1993). Taking a Devonian age of the greywackes in the BLZ as certain, other authors favour the model that the greywackes represent a clastic mixture of detritus from the CSGC and a Devonian arc and that the BLZ is a major suture zone (Loeschke et al., 1998; Hann & Sawatzki, 2000). This interpretation is based on the presence of chrome-spinel detritus in the greywackes, on the Nd isotopic composition of the greywackes and on Middle Devonian ages of detrital, magmatic zircons (Gruler et al., 1999). The sedimentary precursors of the northern, metamorphic part of the BLZ were most likely deposited in the Middle to Upper Silurian (Wenlock-Ludlow: 430-415 Ma B.P.), as indicated by chitinozoans (Montenari & Maass, 1996), but contain also Ordovician chitinozoans (Hann et al., 1995; Montenari & Maass, 1996).

d. THE SOUTHERN SCHWARZWALD GNEISS COMPLEX

Metamorphic rocks

The outcrop level in the Southern Schwarzwald is clearly dominated by granites (Figs. 1, 4, 9).

Two main metamorphic complexes have been recognised in the SSGC: a gneiss-leptynite association and the Wehra-Wiesetal diatexite (Wimmenauer, 1984), mainly described as a 'granitised' biotite (-hornblende) gneiss with lenses of amphibolites, metagabbros and metaanorthosites (Metz & Rein, 1958; Hann & Sawatzki 1998). Whereas the former authors favour a model of in-situ granitisation of the Wehra-Wiesetal diatexite, the latter describe a nappe structure with the Wehra-Wiesetal diatexite thrust over the gneiss-leptynite association. The nappe model is mainly based on the fact that granitic and aplitic dikes crosscutting the Wehra-Wiesetal diatexite can not be found in the gneiss-leptynite association (Hann & Sawatzki, 2000). Chen et al. (1999) report U-Pb and Pb zircon evaporation ages of 347-349 Ma for 'anatectic gneisses' of the Wehra-Wiesetal diatexite and of 341-344 Ma for granitic dikes.

It is not yet clear whether or not CSGC and SSGC represent distinct crustal segments with different tectonometamorphic evolutions before Variscan collision. Arguments named in favour of a separate evolution (Stenger et al., 1989) are the lack of a strongly laminated lower crust and a low-velocity zone beneath the SSGC as opposed to the CSGC, the absence of eclogites, garnet-bearing peridotites and orthogneisses in the SSGC as opposed to the CSGC, the lack of metagabbros and metaanorthosites in the CSGC as opposed to the SSGC and the K-dominated composition of the leptynites in the SSGC as opposed to the Na-rich character of those in the CSGC. Loeschke et al. (1998) and Hann & Sawatzki (2000) interpreted SSGC and CSGC as separate terranes and the BLZ as a major suture zone (see above). Arguments in favour of a more differentiated view are that the structure of the lower crust and mantle observed today (e.g. Lüschen et al., 1987; Eisbacher et al., 1989; Edel & Fluck, 1989) is not necessarily that present during Variscan times, that both the CSGC and the SSGC are composite metamorphic complexes and need not necessarily contain exactly the same rocks and that there are quite some striking similarities between the leptynite-gneiss complex of the SSGC and units 2 and 3 of the CSGC (see below).

Based on lithology and metamorphic grade, the rocks of the leptynite-gneiss complex can be divided into a variegated unit devoid of high-pressure relics (corresponding to unit 2 in the CSGC) and a gneiss unit with granulite-facies relics (corresponding to unit 3 in the CSGC; Hanel & Wimmenauer, 1990; Fig. 4). Cordierite-sillimanite gneisses and migmatites in the variegated unit devoid of high-pressure relics record a HT-LP metamorphic stage at the same conditions and at the same time as the unique HT-LP overprint in the CSGC (Kalt et al., 1994a; Kalt, 1995, Hautmann et al., 1996). The granulites in the SSGC are mainly of acid (quartz + ternary feldspar ± garnet ± sillimanite) to intermediate (quartz + ternary feldspar ± garnet ± orthopyroxene ± clinopyroxene) to basic (orthopyroxene + clinopyroxene + plagioclase ± garnet) composition. Metapelitic granulites as those of the CSGC are absent. However, the equilibration conditions calculated for the granulite-facies stage in the SSGC granulites (1000-1050°C / 1.3-1.7 GPa; Kalt et al., 1999b) are very close to those of unit 3 in the CSGC (950-1000 °C at minimum pressures of 1.5 GPa; Marschall et al., 2000). Moreover, the granulites of the SSGC contain relics of spinel peridotite and garnet-spinel peridotite (Kalt & Altherr, 1996; and unpublished data) as those of the CSGC. Equivalents of unit 1 and hence relics of eclogites have not been found in the SSGC.

The remnants of layered intrusion (metagabbros, metaanorthosites, metaperidotites) occur in a discontinuous belt of 23 km length mainly in the Wiesetal-Wehratal area (Sebert & Wimmenauer, 1992). The metabasites form isolated xenoliths of a few decimetres up to 100 m size. A pronounced compositional layering, obtained during magmatic cumulate formation, and a foliation parallel to layering are common features of the rocks. Primary magmatic minerals (olivine, clinopyroxene, orthopyroxene, plagioclase) are partly replaced by corona textures and amphibole. From these features, a two-stage metamorphic evolution in the granulite facies and in the amphibolite facies was deduced (Sebert & Wimmenauer, 1992). Pb zircon evaporation ages of 349-350 Ma, obtained from anorthositic cumulates and amphibolites, are interpreted as the age of basic magmatism 'during the final stage of plate convergence' (Chen et al., 1999).

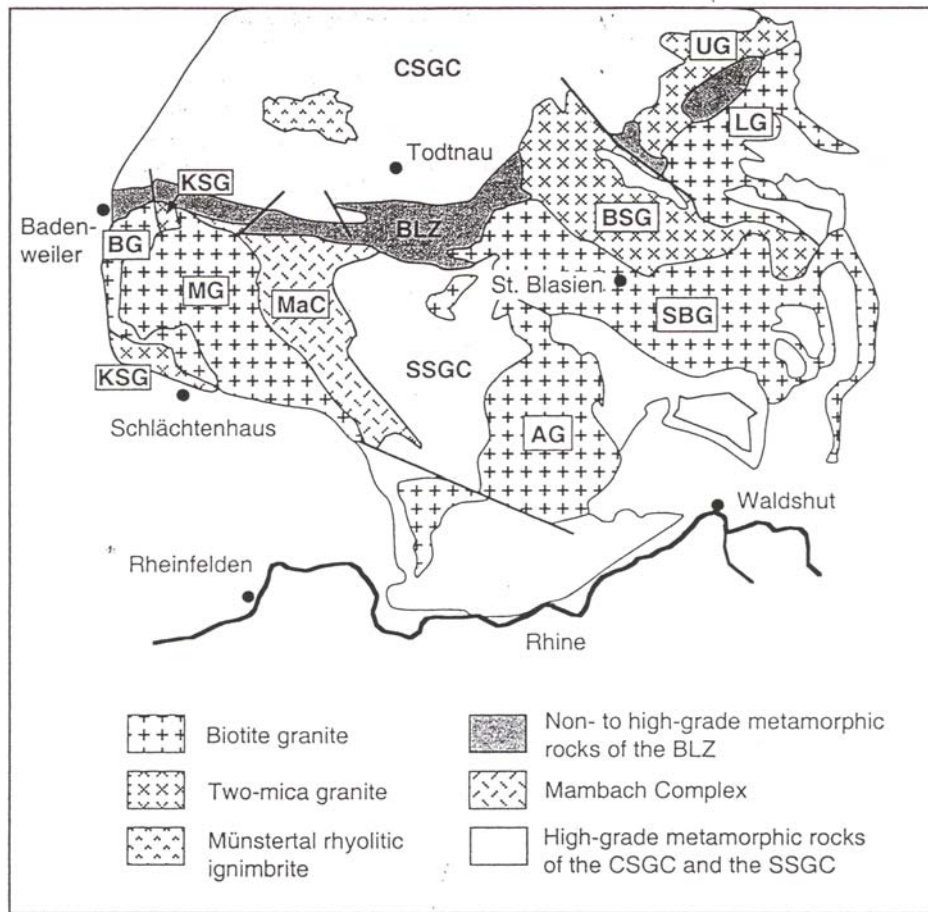


Fig. 9. Simplified geological map of the Southern Schwarzwald Gneiss Complex with its numerous granitic intrusions, modified after Metz (1971). Abbreviations are AG: Albtal granite, BG: Blauen granite, BSG: Bärhalde-Schluchsee granite, BLZ: Badenweiler-Lenzkirch Zone, CSGC: Central Schwarzwald Gneiss Complex, KSG: Klemmbach-Schlachtenhaus granite, LG: Lenzkirch-Steina granite, MG: Malsburg granite, MaC: Mambach complex, SBG: St. Blasien granite, SSGC: Southern Schwarzwald Gneiss Complex, UG: Ursee granite.

Granites

The granites of the SSGC can be divided into two compositional groups (Fig. 9; Schröder, 1929; Rein, 1961; Hahn-Weinheimer & Johanning, 1963; Hahn-Weinheimer & Ackermann, 1967; Emmermann, 1968, 1969, 1970, 1977; Propach & Doppel, 1972; Emmermann & Rein, 1973; Lechner & Hahn-Weinheimer, 1974; Emmermann et al., 1975; Müller-Sohnius et al., 1976; Sawatzki, 1988, 1992): (1) metaluminous to slightly peraluminous biotite granites (Blauen, Malsburg, Albtal, St. Blasien, Lenzkirch-Steina) and (2) peraluminous two-mica granites (Klemmbach-Schlachtenhaus, Bärhalde-Schluchsee, Ursee, Hochfirst, Hammereisenbach). Similar granite types were found underneath the Permo-Carboniferous trough of Northern Switzerland (Mazurek & Peters, 1992; Schaltegger, 1997). Some of the biotite granites from the SSGC (Albtal, St. Blasien) contain abundant mafic microgranular enclaves and rare xenoliths from the country rocks (Otto & Wimmenauer, 1973; Stenger, 1979; Büsch & Otto, 1980). The two-mica granites are hydrothermally altered (e.g. Wendt et al., 1970; Murad, 1974; Hoefs & Emmermann, 1983). The metaluminous to moderately peraluminous biotite granites of the Southern Schwarzwald show a limited range in their initial isotopic compositions with $^{87}\text{Sr}/^{86}\text{Sr}(\text{I})$ and $\epsilon\text{Nd}(\text{I})$ ranging from 0.70816 to 0.70913 and from -4.1 to -5.1, respectively. The strongly peraluminous two-mica granites are characterized by somewhat higher ratios of $^{87}\text{Sr}/^{86}\text{Sr}(\text{I})$ (0.70955 to 0.71097) and lower values of $\epsilon\text{Nd}(\text{I})$ (-5.8 to 5.9) (Fig. 10; Altherr

et al., unpubl. data).

Granitic plutonism in the Southern Schwarzwald (Fig. 9) was contemporaneous with or post-dated the accretionary processes documented in the BLZ (Emmermann 1977; Werling & Altherr 1987; Krohe & Eisbacher, 1988). The intrusion age of the foliated Schlachtenhaus two-mica granite is constrained by a $^{207}\text{Pb}/^{235}\text{U}$ age of 334 ± 2 Ma on monazite (Schaltegger, 2000). For the sheared Lenzkirch-Steina biotite granite a single zircon Pb evaporation age of 334 ± 5 Ma was determined (Kober, unpubl. data). Two less deformed biotite granite plutons yielded single zircon evaporation ages of 333 ± 8 Ma (St. Blasien) and 329 ± 4 Ma (Albtal) (Kober, unpubl. data). For zircon fractions from the undeformed Malsburg biotite granite, Todt, (1976) reported a Pb/Pb age of 328 ± 6 Ma (upper intercept). The emplacement of the undeformed Bärhalde-Schluchsee two-mica granite that cuts the BLZ is constrained by a conventional $^{207}\text{Pb}/^{206}\text{Pb}$ age of 332 ± 3 Ma (zircon fractions) and a $^{207}\text{Pb}/^{235}\text{U}$ age of 330.5 ± 1.6 Ma on monazite (Schaltegger, 2000). Nearly concordant and concordant zircon fractions from a granite porphyry dyke at Prag yielded mean $^{207}\text{Pb}/^{206}\text{Pb}$ and $^{206}\text{Pb}/^{238}\text{U}$ ages of 332 ± 2 and 328 ± 2 Ma, respectively (Schaltegger, 2000). The Münstertal (Scharfenstein) rhyolitic ignimbrite (Fig. 9) was deposited as caldera filling on a Viséan/Namurian erosional surface (Maus, 1967; Westphal & Keller, 1995). For this volcanic unit, a conventional U-Pb zircon age of 333 ± 3 Ma was given by Schaltegger (2000). This date is in contradiction to the 325 ± 3 Ma K-Ar and Rb-Sr ages on muscovite and biotite from the Albtal and Bärhalde Schluchsee granites (Wendt et al., 1970, 1974; Schuler & Steiger, 1978). Most likely, the ignimbrite contains zircons inherited from the basement.

2. GEODYNAMIC EVOLUTION

A time-process table is shown in Fig. 10. The key is the variscan orogeny which happened in several stages and lasted from the upper Devonian to the uppermost Carboniferous. It formed a large mountain chain covering an area from today's Alps to southern England

a. Metamorphic evolution

The oldest age information in the Schwarzwald is preserved in the form of Achaean (2.5-2.9 Ga), Early Proterozoic (1.8-2.0 Ga) and 'Panafrican' (540-630 Ma) magmatic cores of detrital zircons from paragneisses of the CSGC. As magmatic events at these times are well known from cratons and mobile belts of former Gondwana, particularly from the African cratons (Gebauer et al., 1989; Nance & Murphy, 1994), a derivation of the oldest components of the CSGC from this area is likely. Ages of 460-520 Ma, obtained on magmatic zircons or zircon domains from paragneisses of units 2 and 3 and on orthogneisses of the CSGC as well as Rb-Sr whole rock data on migmatites of unit 1 (Hofmann & Kohler, 1974) suggest magmatic events and the formation of considerable amounts of new crust at this time.

The entire high-grade metamorphic Moldanubian zone is commonly assigned to the terrane assemblage Armorica on the basis of paleomagnetic evidence from Paleozoic sediments in the Bohemian and the Armorican Massifs (e.g. Tait et al. 1997). Hence, magmatism at 460-520 Ma is most likely related to the break off of Armorica and other terranes from the northern margin of Gondwana in the Ordovician. Sedimentation in marine basins at this time and during the Silurian is documented by chitinozoans and acritarchs for parts of the Moldanubian CSGC and BLZ. Sedimentation in parts of the Saxothuringian BBZ is earlier (Upper Cambrian to Lower Ordovician). The paleogeographic position of the SSGC where no SHRIMP investigations on detrital zircons have so far been carried out is as yet unclear as is. The youngest detrital magmatic zircons in the CSGC were found in metapelitic granulites of unit 3 (391-399 Ma), giving a maximum age estimate for sedimentation and for metamorphism in this unit. Similar ages were obtained on detrital magmatic zircons from the BLZ (371-384 Ma; Gruler et al., 1999). No Silurian or Devonian ages are found in the other units.

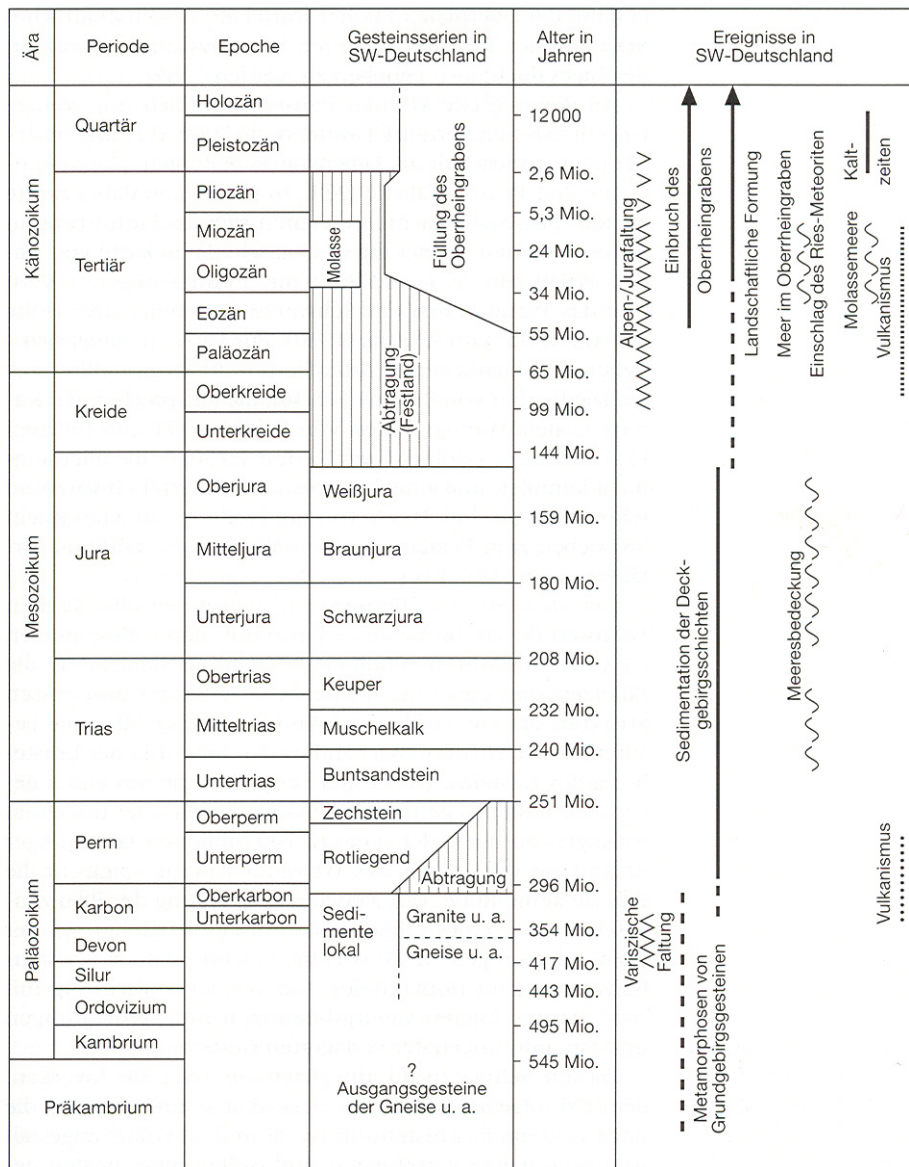


Abb. 10: Geologische Zeittafel mit den wichtigsten geologischen Prozessen im Schwarzwald

The metamorphic evolution before the Carboniferous HT-LP overprint in the CSGC and the SSGC is not yet resolved. Some authors argue in favour of an Ordovician metamorphic event (approximately 480-500 Ma) on the basis of Rb-Sr whole rock data (Hotmann & Kohler, 1974), conventional U-Pb zircon data (Steiger et al., 1973) and Pb evaporation zircon data (Kober, 1986; 1987) on migmatites and gneisses, on the basis of Pb-Pb dating of garnet, K-feldspar and whole rock samples (Chen et al., 1998b) and on the basis of the age and isotopic composition of the orthogneisses of the CSGC, interpreted as indicating a Cambrian magmatic arc (Chen et al., 2000). However, the applied dating techniques are not suited to distinguish magmatic from metamorphic events. In the CSGC, the highest ages unequivocally related to metamorphism are found in isolated relics of eclogite in unit I, in granulites of unit 3 and in the kinzigite. Dating of zircon domains and Sm-Nd isochrons suggest that eclogite-facies metamorphism of probably Ordovician basaltic rocks occurred around 343 Ma, most likely related to subduction of oceanic crust. The entire unit 3 (Silurian/Devonian sediments and magmatic rocks) was subjected to granulite-facies conditions (approximately 1000 °C / 1.5 GPa) in the lower crust around 335-340 Ma ago through burial and heating, as constrained by dating of zircon. Unit 2 (marine sediments deposited after 460-490 Ma B.P.) experienced a MP-MT metamorphic stage before the Carboniferous HT-LP overprint, but its age is unclear. No metamorphic stage prior to the HT-LP conditions is known from unit 1.

At approximately 330-335 Ma, units 1, 2 and 3 of the CSGC and the SSGC experienced HT-LP

metamorphism in the mid to upper crust. HT metamorphism in the lower and middle to upper crust in the CSGC and the SSGC occurred in a compressive setting as indicated by large-scale structures (Krohe & Eisbacher, 1988; Flottmann, 1988; Flöttmann & Kleinschmidt, 1989; Echtler & Chauvet, 1991/92). The peak pressures calculated for unit 3 indicate that parts of the crust were thickened to at least 50 km. Temperatures calculated for units 1 and 3 indicate an extremely high heat flow into the crust. These requirements are all met in a continental arc setting, where subduction and dehydration of oceanic lithosphere triggers melting in the asthenospheric mantle wedge and the generation of large volumes of mafic melts. The HT units 1 and 3 were then (prior to approximately 330 Ma) thrust onto unit 2 which did obviously not equilibrate in a crustal section with high heat flow.

b. Granitoid magmatism

The peak of HT-LP metamorphism is contemporaneous with or slightly older than the formation of zircon in most of the granitic magmas that intruded the metamorphic rocks. The ages of the granites can be used to place time constraints on the thrusting of the CSGC onto the non- to high-grade metamorphic tectonic slivers of the BBZ and the BLZ. The zircon and monazite ages of granites deformed during SE-ward emplacement of the CSGC onto the BLZ (Schlachtenhaus; Lenzkirch-Steina) give a maximum age for thrusting of 334 ± 2 Ma (Schaltegger, 2000; Hess et al., 2000; see above). On the other hand, the U-Pb age of 330.5 ± 1.6 Ma on monazite from the Bärhalde-Schluchsee granite (Schaltegger, 2000) that cuts the thrust zone gives a minimum age for thrusting and marks the change from a compressive to an extensional setting (Echtler & Chauvet, 1991/92). The time of this change in the stress regime is also constrained by U-Pb zircon ages from the Albtal pluton that intruded into a high-angle ductile detachment zone with retrograde stretching lineations marked by sillimanite to chlorite (Echtler & Chauvet, 1991/92). Whereas conventional dating on zircon fractions from this pluton yielded an upper intercept of 334 ± 3 (Schaltegger, 2000), single zircon Pb evaporation gave an age of 329 ± 4 (Kober, unpubl. data).

In the BBZ, a maximum age for the accretion of unit B to the CSGC (possibly represented by group A rocks) is constrained by $40 \text{ Ar}/39 \text{ Ar}$ ages on muscovite (334.5 ± 2.0 Ma) and biotite (331.4 ± 2.0) (Altherr, unpubl. data). On the other hand, a minimum age for the accretion of unit C is given by the intrusion of the Baden-Baden granite (Wickert et al., 1990). Although no age data are as yet available for this granite, an intrusion age of 331 ± 5 Ma may be assumed, since this granite is an equivalent of the younger granites in the Northern Vosges (Altherr et al., 2000), for which an age of 331 ± 5 Ma has been established (Hess et al., 1995). Hence, the age of thrusting of the CSGC northward over the BBZ and southward over the BLZ is restricted to a time span of approximately 4 million years (330-334 Ma B.P.). The Northern Schwarzwald Granite Massif (Fig. 1), however, is significantly younger, as indicated by zircon ages of 325 ± 7 Ma and $40 \text{ Ar}/39 \text{ Ar}$ mica ages of approximately 315-320 Ma (Hess et al., 2000).

Within both the SSGC and the CSGC, granitoid magmatism was immediately followed by the more or less contemporaneous formation of lamprophyric and rhyolitic dikes (e.g. Lippolt et al., 1975; Schleicher, 1978; Lippolt et al., 1983; Sawatzki, 1988; Hradetzky & Siebel, 1990; Kölbl-Ebert, 1992; Hegner et al., 1998). Whereas the lamprophyric magmas were derived from an enriched lithospheric mantle source (Hegner et al., 1998; Altherr, unpubl. data), the rhyolitic dikes (granite porphyry, granophyre) can either be attributed to one or the other dominant granite pluton (e.g. Seebach, Triberg, Malsburg plutons) or represent independent magma pulses (Schleicher, 1978; Sawatzki, 1988).

For the granitoid plutons of the Schwarzwald and the Vosges in general, an origin by multiple intrusion of crustal-derived felsic and mantle-derived mafic (mainly dioritic) magmas was suggested (Altherr et al., 1999a, 2000). In principle, dioritic magmas could also be derived from a hot mafic lower crust instead of originating from the mantle. Numerous experimental studies have shown that at temperatures in excess of $1100 \text{ }^\circ\text{C}$, mafic metaluminous low-silica ($<56 \text{ wt}\%$) melts can be produced by dehydration melting of various metabasaltic compositions (Patino Douce & McCarthy, 1998). Lower, partly mafic crust is represented by the granulitic relics in unit 3, with calculated equilibration temperatures near to $1100 \text{ }^\circ\text{C}$ (see sections on the CSGC and the SSGC). However, the experimentally

obtained melts are generally characterized by low Mg# (<44) and high contents of Na₂O (> 4.3 wt%) and these features are not shown by the diorites from the Schwarzwald and the Vosges (von Drach & Lippolt, 1974; Altherr et al., 1999a, 2000). The origin of these rocks thus remains to be resolved.

The Nd and Sr isotopic signatures of Early Carboniferous granitoid rocks from the Odenwald, Schwarzwald and Vosges are presented in Fig. 11. Compared to the plutons from the Saxothuringian zone, those from the Moldanubian zone have lower values of $\epsilon_{Nd}(I)$ and higher values of $^{87}Sr/^{86}Sr(I)$. Which it is suggested to reflect different source compositions. This hypothesis is substantiated by corresponding differences in some chemical parameters. Compared to the Saxothuringian granitoids, those of the Moldanubian zone are characterized by lower ratios of CaO/(MgO+FeO), K/Rb and Sr/Nd but higher ratios of Rb/Sr, suggesting a more evolved nature of the Moldanubian crust. The compositional difference between the Saxothuringian and Moldanubian lithospheres suggests that the BBZ and its continuation to the west, the Lalaye-Lubine zone in the Vosges, may represent a suture zone between two terranes.

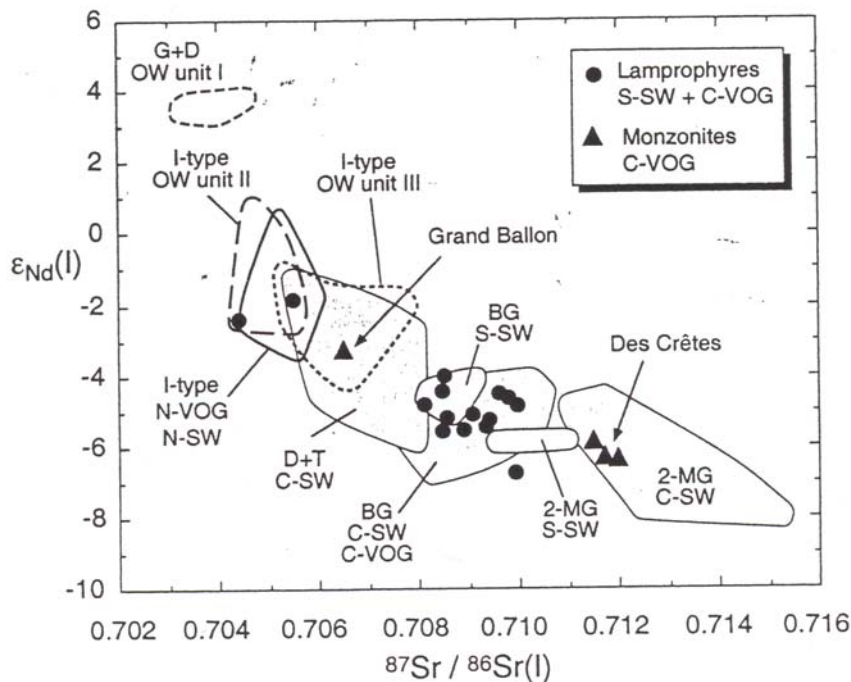


Fig. 11. Initial $\epsilon_{Nd}(I)$ value [$\epsilon_{Nd}(I)$] vs initial Sr isotopic ratio [$^{87}Sr/^{86}Sr(I)$] of lower Carboniferous plutonic rocks from the Odenwald (OW), Schwarzwald (SW) and Vosges (VOG) and of lamprophyres from the Moldanubian parts of the Vosges and Schwarzwald. The data fields for the I-type plutons emplaced into the three tectonic units of the Odenwald (I to III from north to south) are from Altherr et al. (1999b); G = gabbro, D = diorite. Data field for the I-type granitoids (diorites, granodiorites, granites) from the northern Vosges and Schwarzwald (Saxothuringian zone) are from Altherr et al. (2000). Data for granitoids from the Moldanubian parts of the Vosges and Schwarzwald have been further subdivided according to rock type (D = diorite, T = tonalite, BG = biotite granite, 2-MG = two-mica granite) and location (C-SW and C-VOG = central Schwarzwald and central Vosges; S-SW = Southern Schwarzwald to the south of the BLZ). Data for the Moldanubian plutons are taken from Brewer & Lippolt (1974), von Drach et al. (1974), Schuler & Steiger (1978), Liew & Hofmann (1988), Langer et al. (1995), Altherr et al. (1999), Kober et al. (2000), Altherr (unpubl. data). Data sources for lamprophyres: Hegner et al. (1998), Langer et al. (1995), Altherr. (unpubl. data).

C. Hydrothermale Mineralisationen im Schwarzwald

Modifiziert nach: Rudolf Metz, Max Richter und Horst Schürenberg (1957) Die Blei-Zink-Erzgänge des Schwarzwaldes. Beihefte Geologisches Jahrbuch, 29, 277S.

In vielen Teilen des Schwarzwaldes gab und gibt es Erzvorkommen, meist in Gängen, deren erste schon in der Jungsteinzeit genutzt worden sind. Heute ist nur noch die mächtige Grube Clara in Oberwolfach in Betrieb. Dort wird Schwerspat abgebaut, der als Füllstoff für Kunststoffe und Papier und für Beton im Reaktorbau verwendet wird. Außer diesem Bariumsulfat sind mehr als 200 andere Mineralarten gefunden worden. Eine in ihrer Vielfalt und Präsentation begeisterte Ausstellung von besonders schönen Stücken aus dem Schwarzwald überhaupt bietet das Mineralienmuseum in Oberwolfach. Im standfesten Grundgebirge sind zahlreiche Stollen und Grubenbaue aus dem 18. Jh. und dem Mittelalter noch befahrbar.

Der früheste Bergbau vor etwa 5000 Jahren fand nur in den westlichen Randlagen des Gebirges statt, zum Beispiel in der Nähe des heutigen Bad Sulzburg; er ging auf Rötel, der dort als Hämatit in Quarz-Schwerspat-Gängen vorkommt. Auch römischer Bergbau ist gesichert. Der bekannteste Bergbau ging auf die Silber-Blei-Lagerstätten im südlichen und die Silber-Kupfer-Lagerstätten des mittleren Schwarzwaldes. Die Silbererze des Schauinsland waren eine wesentliche Grundlage des Reichtums der Freiburger Bürger im 13. und 14. Jahrhundert; er ermöglichte ihnen den Baubeginn des Münsters. Weiter wurden neben Eisenerzen auch solche mit Kupfer, Blei, Kobalt, Nickel, Wismut und kurzfristig Zink an verschiedenen Stellen im Laufe der Zeiten abgebaut. Dazu kamen Flussspat und Schwerspat, die beide reichlich und weit verbreitet als Hauptgangarten vorkommen. In jüngster Zeit wurden Uran-Vorkommen "interessant". Die einzige Abbaustätte im Krunkelbachtal bei Menzenschwand wurde nach 30-jährigem Untertageabbau 1990 geschlossen.

Vorkommen und Typen von hydrothermalen Mineralisationen

Die Erzgänge, denen der alte Bergbau nachging, sind ungewöhnlich zahlreich. Im Schwarzwald sind nahezu tausend Mineral- und Erzgänge bekannt, die ungleichmäßig verteilt sind. Im Grundgebirge häufen sich die Gänge zwischen Schauinsland, Blauen und Feldberg sowie im Kinziggebiet. Es können Gangreviere unterschieden werden, die entweder eine einheitliche Mineralführung besitzen oder in denen sich Gänge verschiedener Mineralführung überlagern. Im Südschwarzwald überwiegen Blei-Zinkerzgänge, während der mittlere Schwarzwald Gänge der verschiedenen hydrothermalen Gangformationen aufweist. Im Nordschwarzwald veranlassten im Buntsandstein aufsetzende Gänge einen zeitweise lebhaften Silber-Kupfererz- und Eisenerzbergbau. Eine (nicht vollständige) Übersicht über die wichtigsten im Schwarzwald verbreiteten Lagerstättentypen gibt die Tabelle 1.

In einzelnen Teilen des Schwarzwaldes lassen sich die Erzgänge einem Erzbringer zuordnen, wie die Eisenglanz-Schwerspatgänge bei Eisenbach dem Bärhalde-Hochfirst-Eisenbach-Granit oder die Co-Ag-Bi-Gänge des Wittichener Reviers dem Granit von Wittichen. Die große Zahl der Blei-Zink-führenden Mineralgänge im südlichen und mittleren Schwarzwald kann keiner der bekannten Granitintrusionen zugeordnet werden. Die Gänge setzen sowohl im prävaristischen Gneisgrundgebirge auf, als auch in den geringer metamorphen oberdevonisch-unterkarbonischen Sedimenten und in den varistischen Graniten. Die Vererzung wird als mesothermal bis epithermal hochplutonisch angesehen.

Die Blei-Zinkerzgänge

Während im Südschwarzwald Blei-Zinkvererzungen über einen weiten Bereich des Grundgebirges verbreitet sind, treten Mineralgänge mit Pb-Zn-Paragenesen im mittleren Schwarzwald zurück und sind im Grundgebirge des Nordschwarzwaldes unbekannt.

Im Südschwarzwald können die Gänge des Gebietes südlich der Dreisam in Gruppen mit jeweils gleicher bis ähnlicher Mineralführung zusammengefasst werden. Das von SCHÜRENBERG (1950)

im Teufelsgrund und Schindler aufgestellte Schema der Sukzessionen erwies sich als brauchbarer "Schlüssel" für die Gänge im östlich anschließenden Gebiet von Wieden bis Fahl. Hier

Tabelle 1: Übersicht über die wichtigsten Lagerstätten im Schwarzwald

Mineralführung	Nebengestein	Lage
sü = südlicher Schwarzwald; mi = mittlerer; nö = nördlicher		
1. Metamorphe Lagerstätten		
Nickelmagnetkies	Metabasite	Friedrich-August-Grube bei Horbach und Mättle bei Todtmoos
2. Hydrothermale Erz- und Minerallagerstätten im Grundgebirge und im Deckgebirge		
Brauneisen-Schwerspatgänge und Quarz-Flussspatgänge	Buntsandstein	Pforzheim-Neuenbürg (nö)
Cu-Ag-Bi-führende Quarz-Schwerspatgänge	Buntsandstein	Neubulach (nö)
Cu-Ag-Bi-führende Quarz- Schwerspat-Gänge	Buntsandstein und Rotliegendes	Königswart im Murgtal (nö)
Schwerspat-Quarzgänge mit Cu-Ag-Erzen	Buntsandstein und Muschelkalk	Hallwangen, Aach, Domstetten (nö)
Cu-Ag-führende Schwerspat- Quarzgänge	Rotliegendes und Buntsandstein	St. Christophstal, Freudenstadt (nö)
Schwerspat -Brauneisengänge	Buntsandstein	Freudenstadt (nö)
Eisenglanz- Brauneisen - Schwerspatgänge	Grundgebirge	Durbach (mi)
Flußspat-Schwerspa tgänge	Grundgebirge	Hesselbach-Ödsbach (mi)
Cu-führende Quarz- Schwerspatgänge	Grundgebirge	Rippoldsau , Schapbach (mi)
Ag-Cu-führende Quarzgänge	Grundgebirge	Haigerach-Moosbach (mi)
Brauneisen-Schwerspa tgänge	Buntsandstein	Zunsweier- Reichenbach, Lahrer Vorberge (mi)
Schwerspat- Eisenglanz- Brauneisengänge mit Cu-Ag- Erzen und Eisenglanz-Schwerspa tgänge mit Mn-Erzen	Grundgebirge und Buntsandstein	Benauer Berg, Grube Clara (mi)
Eisenglanzgänge	Grundgebirge	Zell am Harmersbach (mi)
Schwerspatgänge mit Co-Ag- Bi-Cu- und U-Erzen und Schwerspat- Flußspatgänge mit Cu-Bi-Erzen	Grundgebirge	Hohberg bei Ippichen (mi)
Ag-führende Quarzgänge	Grundgebirge	Wittichener Revier(Wittichen Heubach, Reinerzau, Alpirsbach, Schiltach) (mi)
Ag-Sb-führende Kalkspat- Schwerspatgänge	Grundgebirge	Revier Haslach (mi)
Schwerspat-Quarzgänge mit Brauneisen	Grundgebirge	Revier Wolfach
Ag-Sb-führende Quarzgänge	Grundgebirge	Mühlenbach (mi)
Schwerspatgänge	Grundgebirge	Adlersbach
Brauneisen - Eisenspat- Schwerspat-Quarzgänge "Eisensteingänge "	Grundgebirge	Gutach (mi)
Eisenglanz-Mn -Erz-führende Schwerspat-Gänge	Grundgebirge	Glottertal und Elztal (mi)
3. Sedimentäre Lagerstätten		
Steinkohlen	Ober-Karbon	Umgebung von Baden-Baden (Varnhalt, Umwegen) (nö) und Revier Diersburg-Berghaupten (mi)
Oolithische Jura-Eisenerze	Dogger	Kahlenberg und Rötelberg bei Ringsheim (mi) und Schönberg (sü)
Bohnerze	Eozäne Tone auf Malm	Kanden, Aggen, Liel (Markgräflerland) (sü) .
Gips	Muschelkalk Keuper Tertiär	Südliche Vorberge, Dinkelberg, Wutachgebiet (sü)

sind im allgemeinen die gleichen Sukzessionen zu finden, wenn sie auch ein allmähliches Abklingen der hydrothermalen Tätigkeit nach Osten hin aufweisen. Komplizierter sind die Verhältnisse im Westen und Nordwesten, im Münstertal und am Schauinsland. Hier liegen teilweise stark abgewandelte Sukzessionen vor, die jedoch bei näherer Untersuchung ebenfalls auf das Schlüsselschema zurückgeführt werden können.

Die Zusammenfassung der Gänge zu Gruppen erfolgte nach Aufstellung von Sukzessionstabellen für jeden Gang, und ausschließlich nach dem Gesichtspunkte gleicher oder angenähert gleicher Mineralfolge, nicht aber nach der regionalen Lage. Hierbei stellte es sich im Gebiete des Untermünstertales, wo sich verschiedene Gruppen oder Typen unterscheiden lassen, die sich regional überschneiden, heraus, dass Gänge mit gleicher Paragenese meist die gleiche Richtung besitzen. Man kann so auch von der Paragenese her ableiten, dass Spalten eines gleichen tektonischen Systems gleiche Öffnungszeiten hatten und die gleiche Mineralzufuhr erhielten. Diese Beobachtung darf nicht für das Obermünstertal und den Schauinsland verallgemeinert werden, wo offensichtlich zwei verschiedene tektonische Systeme gleiche bis ähnliche Mineralführung besitzen. Die Gänge gehören zwar alle der Blei-Silber-Zink-Formation an, doch möge zur Unterscheidung der Gruppen eine detailliertere Namengebung vorgeschlagen werden, welche die Mineralführung kurz zu charakterisieren vermag (Abb. 12).

Hieraus ist zu erkennen, dass, abgesehen von den 7 Gängen der Gruppe D und den nicht eindeutig charakterisierten Gängen der Gruppe F, etwa 150 Gänge einen deutlichen Zusammenhang von Mineralführung und Gangrichtung aufweisen

		A	A	B	C	D	E	F	G	
		West	Ost							
		Ext. Int.	Ext. Int.	Ext. Int.	Ext. Int.	Ext. Int.	Ext. Int.	Ext. Int.	Ext. Int.	
		Brekzien u. Trümer								
I. Vorphase		Calcit u. Schwersp.Neg.								
I. GENERATION	Vorphase Quarz u. Kiese	Quarz I Hornstein								
		Quarz I - x x								
		Magnetkies								
		Pyrit-Markasit I								
		Arsenkies								
	Hauptphase Buntmetallsulfide u. Spätle	Zinkblende I								
		FluBspat I								
		Bleiglantz I								
		Kupferkies I								
		Ag. As u. Sb.-Min.								
		Antimon - Min.								
		Karbonate I (viol. Ca, Fe)								
		Schwerspat I								
		Quarz II - x x								
II. GENERATION	Vorphase Quarz u. Kiese	Quarz II - pseudom.								
		Pyrit-Markasit II								
		Zinkblende II								
	Hauptphase Buntmetallsulfide u. Spätle	FluBspat II								
		Bleiglantz II								
		Kupferkies II								
		Fahlerz II								
		Karbonate II (Fe, Mg, Fe)								
		Schwerspat II								
		Quarz III								
III. GENERATION	Karbonate u. Kiese	Karbonate III (Fe, Mg, Ca, Mg)								
		Bleiglantz III								
		Pyrit-Markasit III								
		Kalkspat III								
		Oxydations - Min.								

Abb. 12: Übersicht über Mineralverteilung und Mineralogie in den Gängen des Schwarzwaldes

- A. Quarz-Flussspatgänge mit Blei-Silber-Zinkerzen im Gebiet Untermünstertal - Wiesental, Typ Schindler. 65 Gänge: $\pm 5^\circ$ /W (11 Gänge 50° /NW).
- B. Quarz-Flussspatgänge mit Bleierzen im Östlichen Südschwarzwald, Typ Rupprechtgang. 15 Gänge $\pm 5/90^\circ$ (1 Gang $40/90^\circ$).
- C. Quarz-Kies-Fahlerzgänge bei St. Ulrich und im Münstertal, Typ Wildsbach. 17 Gänge 135° /NE.
- D. Quarz-Kies-Antimonerzgänge bei St. Ulrich, im Münstertal und bei Sulzburg, Typ Münstergrund. 7 Gänge, kein bevorzugtes Streichen.
- E. Quarz-Schwerspat-Kalkspatgänge mit Blei-Zinkerzen im Schauinslandgebiet, Typ Schauinsland. 13 Gänge, 25-500/NW, (7 Gänge 1000/S.)
- F. Quarz-Schwerspatgänge mit Blei-Zinkerzen (vereinfachter Typ Schauinsland) im Münstertal, Typ Kropbach. 20 Gänge $0/90^\circ$ im Untermünstertal (4 Gänge 90° /N im Obermünstertal).
- G. Quarz-Eisenspat-Schwerspatgänge mit Kupferkies und Zinkblende im Untermünstertal, Typ Riggengbach. 11 Gänge 200° /ESE.
- R. Gänge ohne ausgeprägten Formationscharakter (meist mit Quarz-Schwerspat-Eisenglanz), Typ Tirolergrund. 8 Gänge $135/90^\circ$.

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Further evidence for a Jurassic mineralizing event in central Europe: K–Ar dating of hydrothermal alteration and fluid inclusion systematics in wall rocks of the Käfersteige fluorite vein deposit in the northern Black Forest, Germany

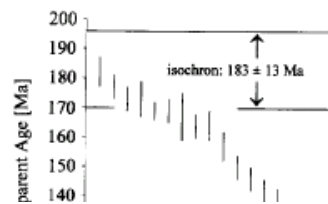
Authigenic sericite record of a fossil geothermal system: the Offenburg trough, central Black Forest, Germany

J. Kalkspat-Kupfererzgang im Rotenbach und

Abstract The hydrothermal fluorite vein deposit of 'Käfersteige' ranks among the biggest in central Europe. It is located along the suture zone that separates the Moldanubian and Saxothuringian units in the northern Black Forest, and is hosted in Bunter sandstone and underlying granitic basement. K–Ar ages of authigenic illite from the wall rocks give a Jurassic formation age of around 145 Ma for the deposit. Age data scatter down to 80 Ma in illite from the clay gouge and indicate a younger Cretaceous-Tertiary hydrothermal overprint. The pyrophyllite component in authigenic illite from wall rocks and the re-equilibration of illite suggest a formation temperature around 200 °C. Secondary fluid inclusions in quartz from the wall rocks define a syn-mineralization fluid episode involving Ca–Mg–K–Na–Cl-rich brines (about 27 wt% NaCl_{eq}) with a T_h of about 125 °C, and a post-mineralization Na–Cl-rich fluid overprint with a T_h of about 100 °C. Both generations of fluid inclusions relate to the final event of each cycle, while authigenic illite composition and re-equilibration of illite in the clay gouge may document initial temperatures of formation. The Upper Jurassic fluid system can be traced all over western Europe. It is probably an expression of continent-wide rifting and concomitant regional fluid circulation in connection with major tectonic disturbances, magmatism and abnormal heat flow during the opening of the North Atlantic ocean. The younger barren fluid overprint could be related to the onset of the Alpine orogeny.

Abstract A fossil geothermal area is hosted by the Carboniferous, Permian and Bunter sandstones of the Offenburg intramontane trough in the central Black Forest. The hydrothermal alteration is identified on the basis of newly formed sericites, which appear as pseudo-morphs after feldspar and filling of pore spaces. According to K–Ar dating of sericite, sericitization occurred about 145 Ma ago (Jurassic). On the basis of $\delta^{18}O$ analyses of sericite, sericite composition and vitrinite reflectance, the hydrothermal fluids had temperatures of 150–210 °C. Because their electrolyte content was low, these fluids are assumed to have derived from meteoric water. A second pulse of electrolyte-rich hydrothermal fluids resulted in quartz overgrowths. Fluid mobilization seems to be linked to the disintegration of Pangaea and to reactivated fault systems extending from the crystalline basement into the intramontane sediments.

Arsenkies-Kobalterz-Gang bei Sulzbach.



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(U + Th)–He evidence of Jurassic continuous hydrothermal activity in the Schwarzwald basement, Germany

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Abstract

(U + Th)–He model and isochron ages of 21 botryoidal hematites collected from the Hohberg base-metal vein system in the Triberg granite complex, Mittelschwarzwald (Germany), cover most of the Jurassic period. The possibility of quantitative ⁴He loss triggered by successive hydrothermal events, although conceivable for individual hematites — based on a ⁴He retentivity study and calculated opening temperatures of hematite for ⁴He of > (90–180)°C — can largely be ruled out. Thus the range of model and isochron ages indicates that the majority of the hematites are not cogenetic. This is supported by a broad range of Pb_{total} concentrations in the Hohberg botryoidal hematites (150–2700 ppm), by comparison with virtually identical Pb_{total} concentrations found in cogenetic hematites from four different vein systems in the Schwarzwald.

The Th/U ratios of the hematites vary by a factor of ~700 (0.01–7.25) mainly as a result of the range of Th concentrations (0.1–74 ppm). The incorporation of substantial amounts of Th in hematite is thought to result from increased Th transport mobility, possibly caused by association with colloids and/or complexing in the ore solution.

$\delta^{18}O$ analyses of a ~160 Ma old cogenetic hematite–quartz pair indicate that precipitation occurred from a fluid of magmatic origin at a temperature of ~180°C. The regional temperature peak in the basement at the present erosional surface during the Jurassic (~130°C) provides a lower temperature limit of hematite formation. This implies that hematite emplacement occurred at temperatures roughly ranging between ~130 and ~180°C. The Jurassic continuous hydrothermal activity suggests that the pre-rifting European craton is characterized by local thermal fluxes during a long time span.

Deep groundwater in the crystalline basement of the Black Forest region

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Two major types of groundwater can be readily distinguished in the Variscian crystalline basement of the Black Forest in S±W Germany. Saline thermal water utilized in spas has its origin in 3±4 km deep reservoirs and developed its composition by 3 component mixing of surface freshwater, saltwater (of ultimately marine origin) and a water±rock reaction component. In contrast to the thermal water, CO₂-rich mineral water, tapped and bottled from many wells in the Black Forest, has low salinities but a TDS distribution similar to that of thermal water. It developed its chemical composition entirely by reaction of CO₂-rich water with the gneissic or granitic aquifer rock matrix. Particularly important is the contribution of various plagioclase dissolution and weathering reactions that may, at some locations, involve precipitation and dissolution of secondary calcite. Sodium/Ca ratios of water and of rock forming plagioclase in the basement rocks suggests that plagioclase weathering is strongly incongruent. Calcium is released to the water, whereas Na remains fixed to the albite feldspar component. The major element composition of 192 water samples used in this study also indicates a clear vertical stratification of the type of water chemistry; Ca±HCO₃ near the surface, Na±Ca±HCO₃±SO₄ at intermediate depth and Na±Ca±Cl at great depth. The mean permeability of Black Forest granite is about $K=10 \cdot 10^{-6}$ m/s; it is significantly lower in gneisses (gneiss: mean $K=5 \cdot 10^{-8}$ m/s) leading to focused flow through granite. Highly permeable fracture and fault zones, particularly in granite, are utilized by high-TDS saline deep groundwater as ascent channels and flow paths. Although spatially closely associated, the topography driven upwelling system of saline deep water and the near-surface flow system of CO₂-rich mineral waters are hydraulically and chemically unconnected.

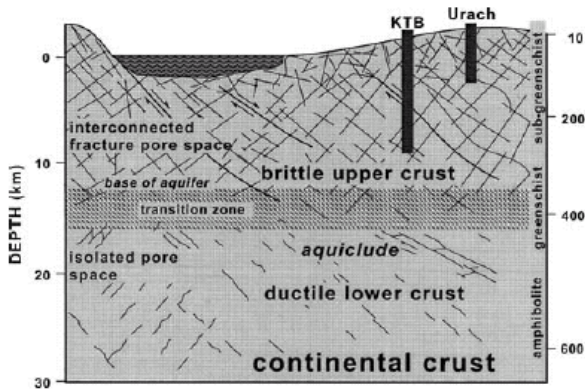


Fig. 1. Schematic cross-section through the continental crust. Brittle upper crust with an interconnected fracture pore space (aquifer) and ductile lower crust characterized by isolated pores (aquiclude) are shown.

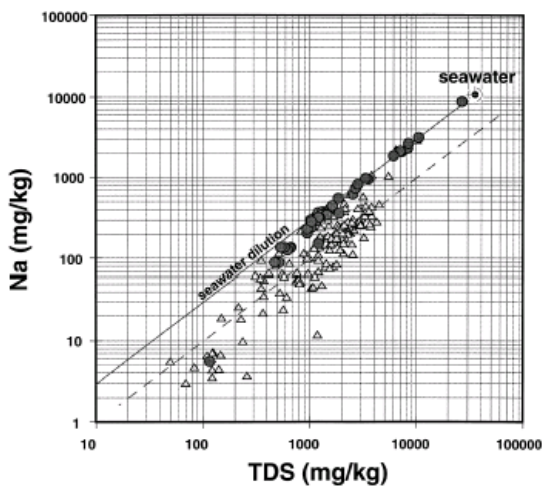
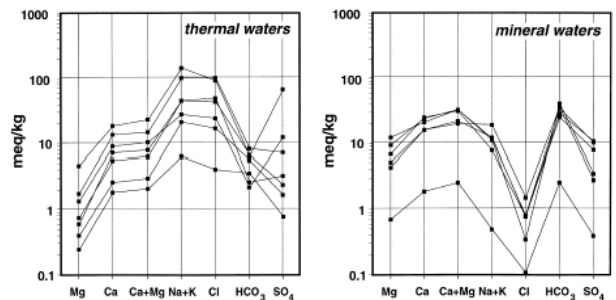


Fig. 6. Na versus TDS; circles: thermal water; triangles: other waters including mineral waters. Thermal water fall close to the sea-water dilution line, mineral waters define a scattered trend that parallels the seawater dilution line at lower Na.

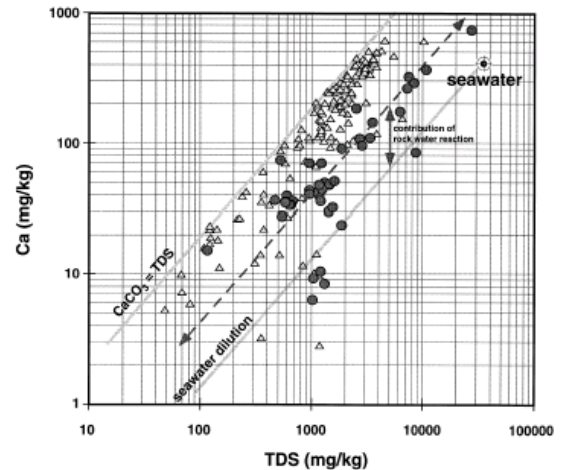


Fig. 8. Ca versus TDS; circles: thermal water; triangles: other waters including mineral waters. In many of the mineral waters Ca is limited by the solubility of calcite. The upper limit is approached if all dissolved solids are exclusively related to calcite dissolution.