Maar-Diatreme Volcanoes, their Formation, and their Setting in Hard-rock or Soft-rock Environments

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ABSTRACT. Maar-diatreme volcanoes mostly form when rising magma in basic to ultrabasic volcanic fields interacts explosively with groundwater. Less commonly, there also exist maars associated with intermediate to acid magmas. The formation and growth of the maar-diatreme volcano type, the second most common volcano type on continents and islands, is reviewed applying the phreatomagmatic model of its formation. The site of explosions is the root zone which penetrates downward on its own feeder dyke. Because of repeatedly developing mass deficiencies in the root zone, the overlying cone-shaped diatreme and the maar crater are the consequent collapse/subsidence features. Prolonged downward penetration of the root zone leads to repeated collapse phases of both the diatreme and maar, thus both grow in size the longer the maar-diatreme volcano is active.

Two contrasting environments exist with respect to groundwater availability for the phreatomagmatic explosions: the hard-rock environment which is a joint aquifer and the soft-rock environment which is a pore aquifer. In the hard-rock environment, the tephra of maar-diatreme volcanoes contains large volumes of rock clasts originally derived from the hard country rocks formerly occupying the root zone resp. the diatreme and maar crater. In the soft-rock environment, the tephra contains large amounts of the individual minerals and pebbles from the sediments but hardly any rock clast consisting of indurated sediments. The two environments are frequently combined in areas where unconsolidated, water-saturated sediments overlie diagenetically indurated sediments and/or crystalline basement rocks or in areas where unconsolidated sediments contain interbedded solidified and jointed sills and lava flows.

Maar-diatreme volcanoes have typically formed in areas characterized by rather normal groundwater conditions. In contrast, areas characterized by highly permeable water-saturated rocks or pebble beds just below the Earth's surface give rise to tuff-rings or tuff-cones. Tuff-cones also form in the shallow sea or in lakes.

KEY WORDS: maar, diatreme, phreatomagmatism, hydrogeology, hard rocks, soft rocks.

Introduction

Occurrence of maar-diatreme volcanoes. After the intermediate, basic and ultrabasic scoria cones, monogenetic maar-diatreme volcanoes are the second most common volcano type on continents and islands (Wohletz and Heiken 1992). The majority of the maar-diatreme volcanoes represent the phreatomagmatic equivalent of the magmatic scoria cones and their associated lava flows (Lorenz 1985, 1986, 1998). Most maar-diatreme volcanoes occur in volcanic fields of basic to ultrabasic composition which comprise several tens to several hundreds of individual monogenetic volcanoes – scoria cones, their lava flows, and maar-diatremes. In addition, maar-diatremes are associated with large polygenetic volcanoes and occur in the foot plains and calderas of shield volcanoes, stratovolcanoes and caldera volcanoes. Finally but more rarely, they also form as a phreatomagmatic equivalent of acidic to intermediate domes.

Scoria cones. Basic to ultrabasic magma in the upper crust mostly flows through dykes of less than, to in excess of, one metre in thickness. Flow direction is either vertical or more or less lateral, as increasingly demonstrated by AMS studies (e.g., Ernst and Baragar 1992). Very close to the Earth's surface, the vertical flow component becomes dominant and has a velocity of several m/s to in excess of 10 m/s. Scoria cones form when magma close to the Earth's surface rises through a dyke and, during its final approach to the surface accelerates because of the formation and growth of vesicles and fragments due to several processes such as hydrodynamic fragmentation in free air, vesiculation in the still fluid state and brittle fragmentation caused by magma flow exceeding the critical shear strength (Zimanowski 1998, Morissey et al. 2000). Associated lava flows either form because of a high production rate resulting in less effective cooling and coalescence on the ground of still fluid clasts or because of quiet effusion. Scoria cones are known to erupt for weeks, months or even years. The scoria cone of Paricutin in Mexico, finally 424 m high, and its associated lava flow field, finally 24,8 km^2 in size, erupted for about 9 years (Feb. 20, 1943 to March 4, 1952; Luhr and Simkin 1993). Also Jorullo in its neighbourhood, which reached a final height of about 350 m and formed a lava field about 1.25 km^2 in size, erupted for about 15 years (1759–1775; Luhr and Simkin 1993). According to Luhr and Simkin (1993) most of the eruptions giving rise to scoria cones are over within one year or less.

Size of maars. The maar-diatreme volcano in principal consists at the surface of the maar crater, which is cut into the pre-eruptive land surface, the tephra ring surrounding the crater, the more or less cone-shaped diatreme, which underlies the maar crater, the irregular-shaped root zone surrounding and underlying the lower end of the diatreme, and finally the narrow feeder dyke at depth (Fig. 1). The maar crater is less than 100 m to over 2 km in diameter (measured from the crest of the tephra ring) and several tens of metres to 300 m deep (measured also from the crest of the tephra ring). The tephra ring is several metres to perhaps over 100 m high. Its inner slope dips towards the interior of the crater at about 33° (natural angle of rest) and the outer slope dips outward at a much shallower angle (e.g., 5–10°), both angles depending on the total volume ejected, the pre-eruptive topography, and, during and immediately after the eruptions of the volcano, also on the state of moisture content of the tephra and erosion processes (e.g., slumping). The tephra ring is built up by possibly a few tens to over of 1000 tephra beds, the majority being only a few mm, cm to 1–2 dm thick. These thinly

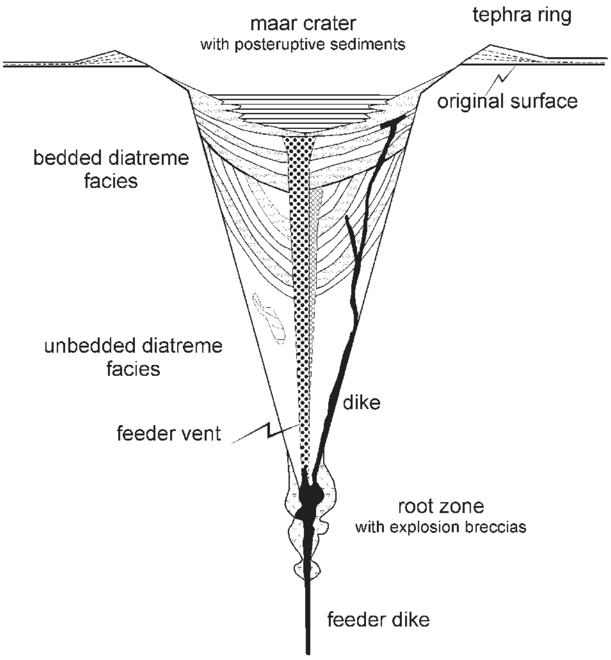

Fig. 1. Schematic diagram of a maar-diatreme volcano showing its feeder dyke, root zone, overlying cone-shaped diatreme (with the lower level showing unbedded volcaniclastics and the upper level showing a saucer-shaped structure with primary pyroclastic beds interbedded with beds derived from reworking of mostly tephra-ring pyroclastic beds), an unconformity in the bedded sequence because of a collapse phase, feeder vents in the centre of the diatreme, the maar crater with its posteruptive background sediments and debris flow and turbidite beds, as well as the proximal tephra ring and the distal tephra veneer. Scale: width equals depth.

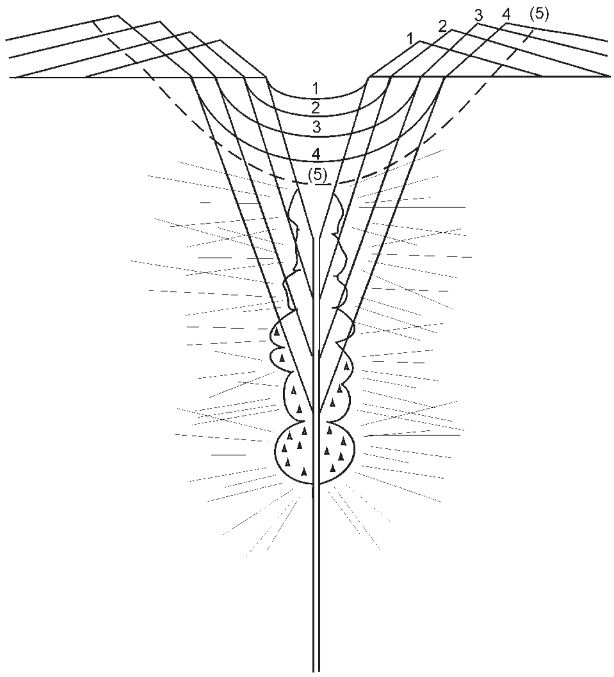

Fig. 2. Schematic growth of a maar-diatreme volcano. During phreatomagmatic explosions the associated shock waves fragment the country rocks in what becomes an explosion chamber and, less intensively, the surrounding country rocks. A number of such explosion chambers form the irregular-shaped root zone.The diatreme and the maar crater grow by downward penetration of the root zone and consequent phases of collapse, resulting in a larger and deeper diatreme and a larger and deeper maar crater. Scale: width equals depth.

bedded tephra beds of the tephra ring represent the proximal tephra deposits and point to the number of eruptions of the maar-diatreme volcano. From the foot of the tephra ring, very thinly bedded tephra extend outwards in a thin veneer for up to a few hundreds of km and represent the distal tephra deposits.

Size of diatremes. Geophysical exploration and drilling as well as investigation of the country-rock clasts in the maar tephra give evidence that maars are underlain by diatremes (e.g., Hawthorne 1975, Lorenz and Büchel 1980, Lorenz 1982a, Büchel 1984, 1987, 1988, Büchel et al. 1987). Exposed diatremes may be several tens of metres to over 1.5 km wide and less than 100 m to over of 2.5 km deep. In hard rocks as is known from diamondiferous kimberlite mines in South Africa they frequently dip inward at average angles of 82° (Hawthorne 1975). The diatreme fill consists of volcaniclastics, subsided blocks of country rocks and a variable amount of intrusive rocks. The volcaniclastics themselves comprise:

1. phreatomagmatic tephra beds in the upper diatreme levels, especially in the larger diatremes, with characteristics as occurring in the tephra ring, and
2. reworked pyroclastics in the upper but more prominently in the lower diatreme levels, derived from the tephra ring and the walls of the maar crater.
3. In addition, the diatreme fill contains phreatomagmatic tephra occupying various vertically orientated channels which represent enlarged original feeder vents through which the tephra clouds were rising to the surface.
4. The diatreme may also contain large blocks and rock slices of country rocks subsided from higher stratigraphic, resp. higher structural levels compared to the country rocks in the diatreme walls.

Historic maars. Despite the frequency of maar-diatreme volcanoes in volcanic fields and other volcanic environments, only very few maar-diatremes formed in historic times. The most recent ones are the Nilahue maar, resp. Carran, which formed in Chile in 1955 (Müller and Weyl 1956, Illies 1959), Iwo Jima, Japan, formed in 1957 (Corwin and Foster 1959), the Ukinrek Maars, which formed in Alaska in 1977 (Kienle et al. 1980, Self et al. 1980, Büchel and Lorenz 1993, Ort et al. 2000) and the Westdahl maar which formed on the Aleutian Islands in 1978 (Wood and Kienle 1990). The maars best studied during and after their eruptions were the Ukinrek Maars (Kienle et al. 1980, Self et al. 1980, Büchel and Lorenz 1993, Ort et al. 2000). The Ukinrek West Maar formed within 3 days; its crater was finally 175 m wide and 35 m deep. Ukinrek East Maar formed in the following 8 days and, after its syneruptive growth, reached a diameter of 340 m and a depth of 70 m. Both craters were filled more (East Maar) or less (West Maar) with a lake. Judging from the comparison of the life span of active scoria cones and maars and their respective sizes, it is highly conceivable that maars with large diameters and high depths of the respective maar crater were active for months or even years and that maars grow in size the longer they are active (Lorenz 1985, 1986).

Formation of maar-diatreme volcanoes

The phreatomagmatic model. The complex process chain in the formation of maar-diatreme volcanoes has been analysed in a number of recent papers (Lorenz 1985, 1986, 1998, 2000a–c, Zimanowski 1986, 1992, 1998, Lorenz et al.1990, 1994, 1999, 2002, Fröhlich et al. 1993, Zimanowski et al. 1995, 1997a–c, Büttner and Zimanowski 1998, Ort et al. 2000, Lorenz and Kurszlaukis in press) and thus will be reviewed only briefly here.

In principal, the maar-diatreme volcanoes form when magma, irrespective of its chemistry (Lorenz et al. 1994), rises through the feeder dyke and, close to surface, interacts explosively with groundwater. The interacting magma–groundwater volume leads to brittle fragmentation of the involved magma volume and the consequent shock waves generated by these thermohydraulic explosions have the quality to fragment the surrounding country rocks (Zimanowski et al. 1997c, Kurszlaukis et al. 1998, Zimanowski 1998, Lorenz et al. 1999, 2000, 2002, Lorenz and Zimanowski 2000). The generation of water vapour from the interacting groundwater leads to further fragmentation of the magma surrounding the interacting magma–groundwater volume and to the rise towards the surface of the eruption cloud thus generated. At the surface, further decompression of the eruption cloud towards ambient pressure and consequent condensation of large amounts of water vapour lead to fallout of the dominant part of the tephra and its deposition on the crater floor and on the surrounding surface by base surges, ballistic transport and some minor tephra fall. Depending on the pre-eruptive topography and interstitial water derived from the phreatomagmatic eruptions and/or from rainfall, the tephra-ring deposits may collapse during crater growth and form lahars, thus thicker accumulations on the crater floor. If deposited on steep relief outside the crater, these deposits may also form lahars in nearby valleys or in other depressions. The distal tephra deposits in the near vicinity of the tephra ring may still contain base-surge material and some ballistics but going outwards, depending on the wind activity, ash falls (and rapidly decreasing lapilli falls) from eruption clouds drifting away from the maar crater rapidly dominate the distal deposits. These distal deposits may extend up to several hundreds of km away from the maar crater.

The primary thinly bedded base surge, ballistic and tephra-fall deposits inside the diatreme are interbedded with thick tephra beds that represent redeposited tephra, i.e. sediments, volcanogenic debris flows and mudflows. These lahars are derived from the moist thinly bedded pyroclastic beds by syneruptive collapse of arcuate slices of the tephra ring onto the inner crater walls and consequent flow onto the floor of this depocentre.

The actual thermohydraulic explosions occur at the top end of the feeder dyke and result in a near-spherical fragmented space, the so-called explosion chamber (Fig. 1 and 2; Lorenz 2000a, c, Lorenz and Zimanowski 2000, Lorenz and Kurszlaukis in press). Partial evacuation of the clasts and fragmented dyke magma from the region of the explosion chamber by the rising eruption cloud leads to a mass deficiency above the feeder dyke. Downward penetration of the explosion site (Lorenz 1985, 1986, 1998) leads to a series of interconnected explosion chambers one below another and in part also laterally next to each other (following mostly the trend of the feeder dyke) and thus to the irregularly shaped root zone (Fig. 2). The total mass deficit in this root zone ultimately leads to increasing rock mechanical instability of the overlying country rocks and the diatreme. After this instability exceeds a critical value, a collapse of the overlying rock volume into the root zone takes place in order to compensate the mass deficit. The collapsing material – in response to the subsiding volume of rocks overlying the root zone – represents a cone of subsidence like a sinkhole; i.e., the diatreme previously overlying the root zone top end, with its lowermost part reaching into the root zone, collapses further down into the root zone and by this process engulfs the top part of the root zone (Lorenz and Kurszlaukis in press). The root zone consequently looses its former upper levels. Since the zone of subsidence propagates upward towards the Earth's surface, the Earth's surface also has to subside and a crater of subsidence origin is consequently formed: the maar crater (Fig. 2). Renewed explosions, further downward penetration of the root zone and more eruptions result in a subsequent phase of increasing mass deficit and collapse. As long as magma–groundwater interaction continues, these repeated collapse phases result in repeated growth phases of the conically shaped diatreme and also in repeated growth of the maar crater (Fig. 2).

From the first collapse phase onwards the maar crater represents a depocentre collecting crater-wall debris falling and flowing onto the crater floor. As long as the maar crater is rather small, the growth of the crater results in undercutting of crater-walls and of the inner parts of the overlying tephra ring. Collapsing country rocks, i.e. rock falls, rock slides, scree, and lahars from collapsing arcuate slices of the tephra ring will collect on the crater floor. From a certain size of the maar crater onwards the crater floor will also represent a depocentre for pyroclastic deposits, i.e., base surge and impact as well as a few tephra fall deposits. Thus primary tephra beds, country rock debris of various kind, and reworked tephra will form alternating beds on the crater floor and jointly subside towards depth in the growing diatreme and consequently they will get overlain by further primary pyroclastic deposits and sediments derived by collapse of the tephra-ring deposits and underlying country rocks. Note that the site where bedded tephra occur now in a diatreme is the site reached after subsidence and not the site where they had been deposited originally.

The magmatic model. It has to be stated, however, that – in contrast to this author's model – a number of authors favour a magmatic model especially for the formation of kimberlite and carbonatite maars and diatremes and relate their formation to very volatile-rich magmas and aspects of explosive exsolution of these volatiles, respectively to intensive fluidization processes (Clement 1982, Clement and Reid 1989, Kilham et al. 1998, Kirkley et al. 1998, Field and Scott Smith 1999, Scott Smith 1999). The individual magmatic models vary quite remarkably. Nevertheless, it has to be realized that, regardless of the model on the formation of these maar-diatreme volcanoes, any model has to comply with the laws of physics and should be supported by respective experiments. With respect to the phreatomagmatic model, experiments show that kimberlite or carbonatite melts can interact explosively with water (Kurszlaukis et al. 1998, Lorenz et al. 1999). Experiments supporting the magmatic model for the formation of maars and diatremes have not been performed yet.

Hard-rock and soft-rock environments of maar-diatreme volcanoes

Two principally different environments exist with respect to the source of groundwater interacting with the dyke magma: groundwater in hard-rock environments and groundwater in soft-rock environments. Several combinations of these environments are possible as will be discussed below.

Hard-rock environment

Joint aquifers. Hard rocks as such are more or less impermeable with respect to the groundwater flow required for starting and supporting the phreatomagmatic explosive activity of maar-diatreme volcanoes at the level of the top end of the respective feeder dyke. Hard rocks, however, are cut by joints and faults, many of which are hydraulically active. Thus hard rocks represent the so-called joint aquifers. Hydraulic activity of these zones of structural weakness varies in respect of their orientation relative to the respective stress field (Lorenz 1973, Lorenz and Büchel 1980, Büchel 1984, 1993). In regions where hard rocks are uplifted, the zones of structural weakness are more hydraulically active and more readily eroded into valleys (Lorenz 1973, 1982b, Lorenz and Büchel 1980, Büchel 1984). This is not to suggest that joints and faults outside the valley floors are not hydraulically active, but they are less so than those underlying valley floors. The particular zones of structural weakness that result in valley formation are used preferentially by water from many ordinary springs and from thermal, mineral, and CO_2 springs on their way to the surface. Hydrogeology, balneology and economic use of CO_2 make use of this relationship. In karstic areas like the Swabian Alb in southern Germany, most sinkholes/dolines are located in valley floors also pointing to preferential hydraulic activity beneath the floors of former water courses (Lorenz 1982a, Lutz et al. 2000).

The classic maar region of the world, the West Eifel, is underlain by Devonian shales, slates, sandstones, greywackes, limestones, dolomites, and Lower Triassic sandstones, thus hard rocks of various kind. Because of the Tertiary and especially Quaternary uplift, the Eifel is highly dissected by more or less deep valleys (Illies et al. 1979, Fuchs et al. 1983). Especially in areas underlain by Lower Devonian rocks there is not enough groundwater supply available for the local communities in the area. Nevertheless, out of the c. 270 (except for 3 phonolites) ultrabasic alkalibasaltic to foiditic monogenetic volcanoes of the Quaternary West Eifel Volcanic Field with the exception of 3 maars all other 67 maars are cut into valley floors, many even at the head of a valley, no matter if the eruptive fissure at depth followed the respective valley trend or cut across the valley. The available exposures in most of these maars show that phreatomagmatic eruptions were active in these maar-diatreme volcanoes from the beginning to the end of their activity; at least this is what the tephra rings preserved from erosion indicate. It is very rare that magmatic eruptions were active between the phreatomagmatic eruptions (Lorenz and Zimanowski 2000). This relationship in the groundwater-poor West Eifel also points to the fact that groundwater was available underneath the valley floors but it also demonstrates that the structural zones of weakness lying underneath the respective valleys were hydraulically active enough to support the phreatomagmatic activity of the maar-diatreme volcanoes from the very beginning to the very end of the local eruptive activity.

Other areas where hard rocks form the country rocks and where maar-diatreme volcanoes are frequently localized in valleys are, e.g., volcanic fields in the Massif Central in France (Camus 1975), the Swabian Alb in southern Germany (Fig. 3; Lorenz 1982a, Keller et al. 1990, Lutz et al. 2000), and the kimberlites in Lesotho, southern Africa (Nixon 1973, 1995). Despite the fact that the valleys in their present depth and shape are younger than the diatremes, the present valleys are successor valleys of earlier valleys cut down on hydraulically active zones of structural weakness due to regional uplift. Thus if the present valleys are the result of erosion on hydraulically active zones of structural weakness then the maar-diatremes at their time of formation in all probability formed on the same zones of structural weakness.

The Ukinrek East Maar erupted in 1977 on the saddle between two shallow little valleys following the trend of a fault zone exposed in the crater walls of the maar (Büchel and Lorenz 1993). And the Nilahue resp. Carran Maar in southern Chile erupted in 1955 in a valley and even collected intermittently surface water from a small stream in its crater (Müller and Weyl 1956, Illies 1959). It has to be stated, however, that there do not have to be valleys in order for maar-diatreme volcanoes to form, it is the availability of groundwater in hydraulically active zones of structural weakness which is required. Nevertheless, it is in uplifted areas that most of these hydraulically active zones of structural weakness get shaped into valleys by erosion.

In many of the Tertiary volcanic fields of Central Europe which formed in diverse hard-rock environments (granitoids, gneisses, schists, basalts, sandstones, limestones, etc.) there existed a great many basic to ultrabasic maars and also initial maars, the latter with either scoria cones or lava lakes (Bussmann and Lorenz 1983, Lorenz 1985, 1998, Keller et al. 1990, Suhr and Goth 1996, 1999, Suhr 2000, Cajz et al. 2000). In the Quaternary volcanic field of the Chaîne des Puys in the Auvergne, maars and initial maars are associated with the basic magmas (Camus

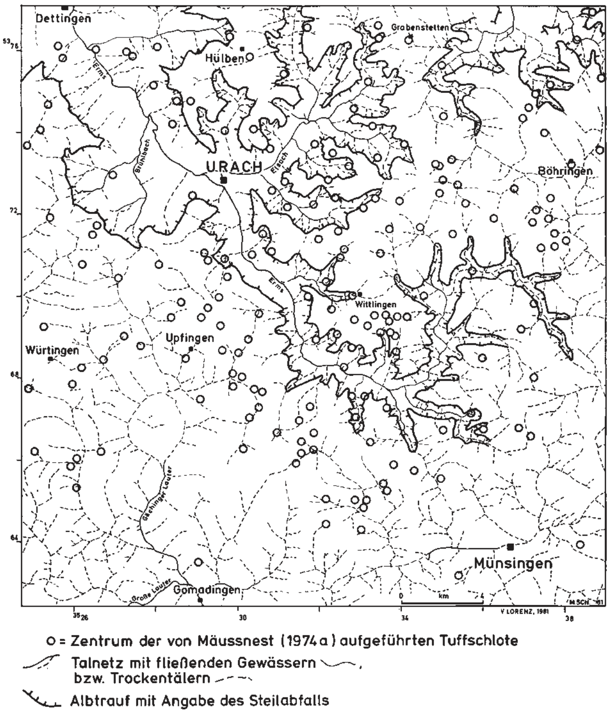

Fig. 3. A diagram showing the relationship between centres of diatremes and valley pattern (perennial streams or dry valleys) in the Swabian Alb. The Swabian Alb is underlain by karstic limestones of Upper Jurassic age. The centres of diatremes are taken from Maeussnest 1974.

1975). In addition, Kilian Crater is a trachytic maar (with a small spine of trachyte in its centre), and a number of the trachytic domes of the Chaîne des Puys became emplaced, in a second, magmatic phase, inside an initial trachytic maar (Camus 1975).

Karstic limestones. Special remark has to be made on karstic limestone environments. The Miocene olivine melilitite Swabian Alb Volcanic Field in southern Germany formed on karstic Upper Jurassic limestones (Fig. 3) which were karstic already in Miocene time (Lorenz 1979, 1982a). Out of about 350 monogenetic volcanoes there formed almost only maar-diatreme volcanoes. The Eisenrüttel in all probability was a lava lake occupying an initial maar (Keller et al. 1990). And the Grabenstetten dyke could have supplied scoria cones and some lava flows (Lorenz 1982a). All other localities represent maar-diatreme volcanoes (Lorenz 1979, 1982a), thus there was so much karst groundwater that the ultrabasic magma had hardly a chance to reach the surface without interacting explosively with groundwater. Other karstic limestone areas which support maar-diatremes are found in the northern Hegau Volcanic Field (Keller et al. 1990), in the Causses, Massif Central, France, and in several kimberlite volcanic provinces in China (Zhang et al. 1989). The diamondiferous Mbuji Mayi kimberlite maar-diatreme volcano in Kasai, Congo, also formed within a sedimentary series containing karstic limestones (Demaiffe et al. 1991).

Late intrusives and extrusives. Magnetic traverses across most maars in the West Eifel (Büchel 1984, 1987, 1988, 1993), however, demonstrate that about 40 % of the maars are associated with local magnetic anomalies of higher intensity than the rest of the near-surface diatreme fill and crater fill. These highs point to local magmatic deposits underneath the present crater floors indicating that, at the end of the phreatomagmatic maar-diatreme activity, magma must have intruded these diatremes to near-surface levels. In many instances magma probably also erupted magmatically onto the former crater floor, the scoria or lava lake being now covered by post-eruptive deposits (Büchel 1984). The gravimetric and magnetic investigation of the Pulvermaar by Diele (2000a, b) also points to a late magmatic activity in that particular crater.

Maar lakes. After the eruptions ended, the maar craters filled with groundwater and surface water because the craters undercut the previous level of the valley floors (Fig. 1). Only the youngest maars still contain a lake, especially when they are large enough and cut off from a potential influx of fluvial sediments. The older and the smaller younger maar-crater lakes are filled by post-eruptive sediments of various kinds and the maar craters ultimately change into "dry" maar craters.

Scoria cones or lava lakes with an initial maar. Zones of structural weakness also exist underneath the valley slopes and on Tertiary and Pleistocene plateau areas between the valleys, as, e.g., in the Eifel (Lorenz and Büchel 1980, Büchel 1984). Judging from the volcanoes formed on these zones they were less hydraulically active than those underneath the valleys. When in the West Eifel ultrabasic magma rose along or cut across hydraulically active zones of structural weakness outside the valleys, this also resulted frequently in phreatomagmatic activity and formation of maar-diatreme volcanoes. At almost all of these localities (with the exception of the three above mentioned maars), an initial maar-forming phreatomagmatic phase was followed by a magmatic phase which produced a scoria cone in the initially formed maar (Fig. 4). This volcano type was called scoria cone with an initial maar by Lorenz and Büchel (1980). Some initial maars are relatively

Fig. 4. Tephra-ring deposits from the Hasenberg volcano in the Quaternary West Eifel Volcanic Field (Lorenz and Büchel 1980). Typical maar tephra rich in Lower Devonian rock clasts (sandstones, slates) formed in an initial phreatomagmatic phase and is overlain by scoria formed by the following magmatic phase. Thus a scoria cone formed within an initial maar. Hammer for scale.

Fig. 5. Typical bedded upper-level pyroclastic deposits from the Rödern diatreme in the Carboniferous-Permian Saar-Nahe Basin, SW Germany (Lorenz 1971, 1972). Juvenile clasts are well visible. The matrix between the juvenile clasts contains many individual mineral grains derived from the Carboniferous-Permian country rocks which at the time of volcanism were water-saturated and unconsolidated. Thus the Carboniferous-Permian sediments at the time of synsedimentary volcanism in the Saar-Nahe Basin represented a typical soft sediment environment. The left tephra became oxidized and the right tephra became reduced during the diagenesis. In the left tephra, imbrication of juvenile clasts indicates deposition by base surges.

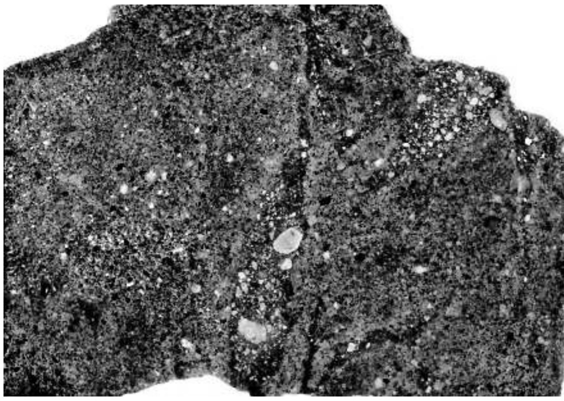

Fig. 6. A block of feldspar-bearing sandstone which syneruptively subsided in the Rödern diatreme in the Carboniferous-Permian Saar-Nahe Basin, SW Germany (Lorenz 1971). The inside of the block contains dykelets and stringers of tephra containing juvenile ash grains and lapilli (bright in colour) showing that the presently indurated sandstone block was not indurated yet at the time the diatreme formed in the Carboniferous-Permian times. Thus, at the time of diatreme formation, the block subsided in the diatreme as a block of unconsolidated sand.

small in size and others are comparatively larger which implies that at the various sites there was relatively shorter or longer availability of groundwater and thus shorter- or longer-lived explosive phreatomagmatic activity prior to a change over to magmatic activity which occurred when magma still kept rising while little or no groundwater remained available. In the West Eifel, outcrop-dependent, out of about 200 scoria cones, approximately two thirds show an initial maar phase. The large number of maars and initial maars (and phreatomagmatic phases within scoria cones) in the West Eifel – and also in the East Eifel Volcanic Field with a smaller number of maars but a large number of initial maars (Schmincke 1977) – points out that the shear availability of groundwater in hydraulically active zones of structural weakness is very conducive to phreatomagmatic explosive activity when magma rises along or intersects such zones of structural weakness.

Root zones in hard-rock aquifers. The root-zone explosion chambers discussed above are formed by shock waves generated by the thermohydraulic explosions. Thus the energy imparted to the hard rocks surrounding the site of explosion produces intensive fragmentation of the surrounding country rocks. Radially outward, the newly generated fractures will decrease in number per unit rock volume but the existing joints and faults may become slightly widened because the shock waves seem to lead to a slight decrease in rock density in the immediate surroundings of diatremes (Diele 2000a, b) implying a slight reorientation of blocks with respect to each other. Opening of other hydraulically active zones of structural weakness in the near vicinity of the main zone of structural activity will give access to new sources of groundwater even when the amount of water is not large. Thus the thermohydraulic explosions create new access of groundwater to the explosion sites (Lorenz 2000a, c, Lorenz and Zimanowski 2000, Lorenz and Kurszlaukis in press). At the maars on the valley floors, the hydraulically active zone of structural weakness underneath the valley floor in all probability will still represent the main conduit for groundwater to the explosions, as at these maars the thermohydraulic explosions are active from the beginning to the end of the eruptive activity. Away from the valleys, however, there is almost always a change over from phreatomagmatic maar eruptions to magmatic scoria production.

Soft-rock environment

A soft-rock environment implies unconsolidated sediments which are water-saturated up to, or close to, the surface. The lack of cement makes all coarser sediments (sands, pebble beds, unconsolidated breccias) a pore aquifer with a high permeability. Syn-sedimentary or post-sedimentary but pre-diagenetic volcanic activity in areas underlain by water-saturated unconsolidated sediments occurs in continental rift zones, molasse basins, late orogenic basin-and-range provinces or coastal sedimentary deposits.

Well-known late orogenic basin-and-range-type grabens exist in southeastern, central, western and southwestern Europe in the late Variscan basin-and-range province (Lorenz and Nicholls 1976, 1984). These grabens contained more or less thick accumulations of water-saturated unconsolidated sediments and most grabens also saw syn-sedimentary volcanism

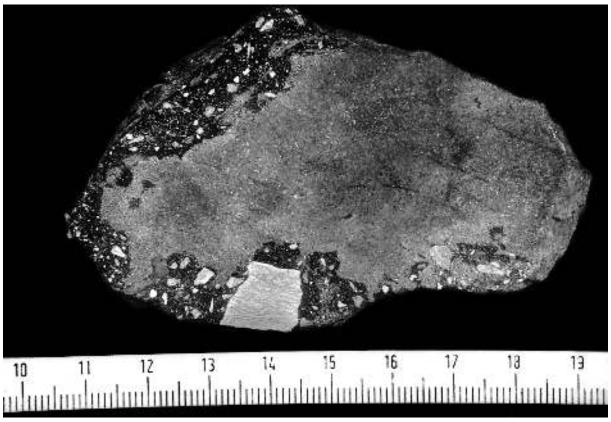

Fig. 7. A block of feldspar-bearing sandstone in the Rödern diatreme in the Carboniferous-Permian Saar-Nahe Basin, SW Germany (Lorenz 1971). Tephra containing juvenile ash grains and lapilli (bright in colour) on the margin of this part of the block, following an irregular serrated contact, shows that the presently indurated sandstone blocks were not indurated yet at the time the diatreme formed in Carboniferous–Permian times.

of acidic in the larger basins also of intermediate to basic silica-saturated magmas. One of the best exposed of these grabens is the Carboniferous-Permian Saar-Nahe Basin in southwestern Germany. The diatremes studied in the Saar-Nahe Basin clearly show that their tephra, in addition to its juvenile ash grains and lapilli (Fig. 5) only contains the individual minerals (quartz, feldspar, mica, clay-minerals) and rock pebbles (milky quartz, quartzite, black chert, rhyolite pebbles) of the surrounding presently indurated sediments. In addition, blocks of sedimentary material like sandstones subsided in the diatremes, and contain dykelets and stringers of juvenile clasts clearly showing that the presently indurated sandstones were still unconsolidated sediments (sands) at the time of subsidence of these blocks inside the diatremes (Figs. 6, 7). From the known downward penetration of the diatremes below their original surface it is evident that at the time of volcanic activity the sediments had not been indurated yet to depths below the syneruptive surface of one to possibly two km (Lorenz 1986, Lorenz and Haneke 2002). The Carboniferous Midland Valley is also a classic area for volcanism having interacted with Carboniferous water-saturated unconsolidated sediments (Francis 1962, 1970, Leys 1982). As in the late Variscan Saar-Nahe Basin, a great many diatremes, containing within their tephra the minerals and pebbles from the surrounding sediments but hardly any indurated sedimentary rock clasts, are associated with sills, laccoliths and intrusive-extrusive domes. A similar environment existed for the volcanism in the Old Red basins in northwestern Europe which represent a late Caledonian basin-and-range-like province. The Basin and Range Province in the western U.S.A. (Heiken 1971), the continental rift zones of the Upper Rhine Graben and the Limagne Graben (de Goer et al. 1998) as well as the main part of the Miocene Hegau Volcanic Field in the Alpine Molasse Basin (Lorenz 1982a, Keller et al. 1990), they all show features related to former explosive interaction of magma with water-saturated unconsolidated sediments.

The Upper Proterozoic diatreme of Argyle, in northern Western Australia formed when diamondiferous lamproite magma interacted explosively with thick water-saturated unconsolidated littoral/coastal sands and muds (Boxer et al. 1989, Stachel et al. 1994).

The diamondiferous kimberlite pipes in northern European Russia are surrounded by very friable and water-saturated sandstones of Vendian (Upper Proterozoic) age (Sinitsin and Grib 1995). At the time of the explosive phreatomagmatic activity, probably in Devonian times, the sandstones were in all probability also unconsolidated and water-saturated. The tephra contains the individual minerals from the sandstones.

The Miocene Ellendale maar-diatreme volcanoes in Western Australia, discussed in more detail in the following chapter, show gently sloping former inner crater walls within unconsolidated sands from the Permian Grant Group. These gently dipping crater walls must have been the result of the soft sediment behaviour of the water-saturated unconsolidated sands which tend to flow when an open crater is formed and thus cannot form walls with steeper angles typical for newly formed maar craters in much more consolidated rocks. The Ellendale maar craters have been filled with younger pyroclastic and reworked deposits from the crater walls as well with lava lakes during the active eruptive period. Maar-diatremes with craters of such low angle walls and steep-walled diatremes are said to be champagne glass-shaped (Smith and Lorenz 1989, Stachel et al. 1994).

Combined hard-rock – soft-rock environments

Hard rocks overlain by soft rocks. A combined hard-rock – soft-rock environment exists when water-saturated unconsolidated sediments overlie jointed hard rocks.

In the Miocene lamproitic Ellendale Volcanic Field, unconsolidated sands of the Permian Grant Group overlie Carboniferous and Devonian jointed shales and limestones (Smith and Lorenz 1989, Stachel et al. 1994). The sands which still today are water-saturated and unconsolidated caused the maar craters to have evolved with very low wall angles and with collapse structures and sand flows towards the crater centre. In contrast, the underlying Carboniferous and Devonian hard rocks (shales and limestones) gave rise to steep-sided diatremes. In almost all of the 48 Ellendale maar-diatreme volcanoes in a second magmatic phase, magma of leucite lamproite or olivine lamproite composition erupted magmatically and formed lava domes (in the case of the more viscous leucite lamproites) or lava lakes (in the case of the less viscous olivine lamproites). The reason for this systematic change from phreatomagmatic to magmatic eruptive activity must have been that during the growth of the maar-diatreme volcanoes, the downward penetration of the diatreme root zones to the levels of Carboniferous and Devonian hard rocks (of known very low water yield) prevented sufficient access of groundwater to the rising magma thus causing a change to magmatic eruptions (Smith and Lorenz 1989, Stachel et al. 1994).

The Cretaceous to Tertiary Kimberlite Pipes in the NW Territories, Canada, formed to a large extent in an area where the crystalline basement rocks were overlain by Cretaceous to Lower Tertiary mudstones, sediments from the Western Interior Seaway and its lacustrine successor (Graham et al. 1998, Nassichuk and Dyck 1998), the youngest mudstones of which were probably still in an unconsolidated water-saturated state during volcanic activity. Likewise the Pliocene nephelinitic Hopi Buttes in Arizona formed in an area where the sediments of Mio-Pleistocene Bidahochi Formation, partly deposited syneruptively, consisted of unconsolidated water-saturated mudrocks (White 1991, 2000) and were underlain by Triassic sandstones and crystalline basement. Again the activity at the many vents started phreatomagmatically, producing maar-diatreme volcanoes, and then terminated when magma intruded the diatremes and rose into the maar craters (White 1991, 2000).

In these volcanic fields in the Northwest Territories and in Arizona the unconsolidated water-saturated muds overlying the jointed hard rocks may have become liquefied during synvolcanic earthquakes and explosive volcanism releasing water to interact with the rising magma.

Soft rocks interbedded with hard rocks. Another combined hard-rock – soft-rock environment is formed when jointed lava flows or sills of synsedimentary volcanic activity are interbedded with unconsolidated water-saturated sediments. Examples of this environment are the Carboniferous-Permian Saar-Nahe Basin in southwest Germany and the Carboniferous Midland Valley in Great Britain. As it has been pointed out above, the unconsolidated sediments penetrated by the maar-diatreme volcanoes appear in the form of individual minerals and individual pebbles. The interbedded hard rocks, i.e., the volcanics of the interbedded lava flows and sills, occur in the pyroclastic rocks of the maar-diatreme tephra as volcanic rock clasts, i.e. accidental lapilli and blocks.

When the diatreme fill and proximal or distal maar tephra of old volcanic fields contain only the individual minerals and pebbles from the neighbouring sediments, it may be argued that the explosive activity caused complete disintegration of the possibly indurated sediments into their mineral and rock clast components. Judging from the situation in clear-cut syneruptive hard-rock environments, where rock clasts are always incorporated in the pyroclastic deposits, and from impact craters in hard-rock environments where rock clasts also always occur in the ejecta rim, no explosion reduces hard rocks to their individual constituents.

Environments with normal and highly permeable aquifers

Normal aquifers. With respect to their hydraulic productivity, both the hard-rock and soft-rock environments are frequently more or less normal aquifers: here, normal groundwater conditions are considered to exist when a well test would result in fast or slow formation of a cone of depression of the groundwater table. Under such normal groundwater conditions, a maar-diatreme volcano can form because only under these conditions does downward penetration of root zones seem possible (Lorenz 1985, 1986, 1998).

Highly permeable aquifers. In contrast, both types of aquifers can be very highly permeable, e.g., in coastal environments where intensively jointed basalt lava flows, coral reefs, or former karstic limestones may be penetrated by the seawater, such as on Oahu where near-coastal rise of magma through such highly permeable hard rocks gave rise to the tuff-rings and tuff-cones of the Honolulu series. Two tuff rings of Hverfjell and Ludent occur in a similar but non-marine environment of intensively jointed basaltic lava flows close to Lake Myvatn (Lorenz 1986). Highly permeable gravel deposits occur along many rivers. Examples of such highly permeable water-saturated gravel beds are the tuff rings of the Menan Buttes at the Snake River, Idaho (Hamilton and Myer 1963, Lorenz et al. 1970, Lorenz 1985, 1986), and many tuff rings along the rivers in central Iceland. The rise of magma into such water-rich soft-rock environment also leads to the formation of tuff-rings which ideally contain only a few percent of clasts of country rock. The highly permeable and water-rich environments do not allow a pronounced downward penetration of the root zone. These phreatomagmatic volcanoes therefore lack a pronounced diatreme formation by repeated collapse into a downward penetrating root zone with a pronounced subsidence crater. A maar crater does not form in this kind of environment but instead a tephra ring or a tephra cone develop, surrounding a wide or small crater above general ground overlying possibly only a small diatreme, i.e., the so-called tuff rings and tuff cones. Even more external water is available in lacustrine and shallow marine environments. A typical example is Surtsey Island which formed off the coast of southern Iceland in the North Atlantic where the sea floor was lying at a depth of 130 m (Thorarinsson et al. 1964). These water-rich environments lead to formation of more wet or moist deposits than in ordinary maar-diatreme environments, i.e., more vesiculated tuffs and accretionary lapilli are formed (Lorenz 1974).

Diagenesis of diatreme fill in hard-rock and soft-rock environments

After the eruptions of a maar-diatreme volcano have ended, the permeable clastic diatreme fill becomes water-saturated and, as indicated above, the crater will also fill with water up to the level of the local or regional groundwater. Then, diagenesis starts being enabled by the large surface area of the clastic diatreme fill of many clasts of country rock, of unstable glass and high-temperature mineral association of the juvenile clasts.

Diagenesis of diatreme fill in hard-rock environment. In hard-rock environments, the diatreme, up to 2–2.5 km deep and of respective diameter, represents a huge volume, emplaced within weeks to years in the consolidated hard rocks but itself consisting of this highly unstable mixture, will undergo an extended diagenetic process chain including hydration, compaction and water escape. Thus for an extended period of time the diatreme fill will go through these diagenetic processes and, because of its conical shape, the fill will subside differentially. Marginal intra-diatreme rocks such as tephra beds will obtain steep inward dips, and internal differential faulting will also take place. Suitable marginal country rocks might get dragged downward. In the case of kimberlites the impressive volume increase due to hydration of olivine to serpentine leads frequently to an overall volume increase of the diatreme fill that ultimately results in some upward movement of the diatreme fill. Suitable country rocks at the diatreme wall, Such as Nama shales in the Gibeon kimberlite province of Namibia, get dragged upwards (Kurszlaukis et al. 1998).

Diagenesis of diatreme fill in soft-rock environment. In soft-rock environments, both the country rock material and the diatreme fill are unconsolidated and water-saturated, the latter at least shortly after the eruptions ended. Compaction and initial diagenetic processes will have started in the unconsolidated country rock sediments much earlier than in the diatreme; its fill formed with the latter having become emplaced rather suddenly and in an appreciable volume. Compaction of the diatreme fill causes differential subsidence and marginal dragdown of suitable country-rock sediments (Francis 1962, 1970, Lorenz 1971). At Argyle, unconsolidated country rocks (mostly quartz sands) transformed into silica-cemented quartzites, and unconsolidated pyroclastic material rich in individual quartz grains in the diatreme also transformed into hard rocks via silica-precipitation in the pore space (Boxer et al. 1989). In the two graben structures of synsedimentary volcanism of the Carboniferous-Permian Saar-Nahe Basin in southwestern Germany and the Carboniferous Midland Valley basin in Scotland, oxidizing and respectively reducing conditions in the graben sedimentary sequence (red beds and drab beds) influenced Eh and pH in the diatreme fill and resulted in reddish or greenish colours of pyroclastic rocks in the diatremes (Fig. 5; Lorenz 1972).

Acknowledgements

I acknowledge the request of Jaromir Ulrych, the Convenor of the Symposium Hibsch 2002 in the Ohre (Eger) Rift region, Czech Republic, to write this paper following the author's attending this symposium.

This manuscript reviews the results of various studies of the present author over more than 35 years and of course also the results of studies of a number of other authors. Fruitful discussions and support by many colleagues and friends as well as financial support by many institutions and companies were therefore too numerous to be named individually, and are gratefully acknowledged. The manuscript was critically reviewed by Felicity E. Lloyd, Reading, Georg Buechel, Jena, and Stephan Kurszlaukis, Johannesburg. Felicity E. Lloyd also considerably improved the English. Their helpful reviews are also gratefully acknowledged.

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