Plateau versus fissure ridge travertines from Quaternary geothermal springs of Italy and Turkey: Interactions and feedbacks between fluid discharge, paleoclimate, and tectonics

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ABSTRACT

Morphologically-different deposits of thermal travertines are known worldwide, but what factors controlled their morphology, volume, and growth for tens of thousands of years is only partially understood. Two main morphotypes of Quaternary thermal travertines are reconsidered here to understand the reasons for their differential growth: the fissure ridge travertines of Denizli Basin, western Turkey, and the travertine plateau of Tivoli, central Italy. For comparable longevities and average vertical deposition rates, the main differences between the studied travertines are as follows: (1) volume of the travertine plateau is about one hundred times larger than each fissure ridge; (2) despite a larger volume, the travertine plateau does not produce relief, whereas the fissure ridges produce a characteristic prominent topography; (3) the travertine plateau grew primarily through lateral progradation, whereas the fissure ridges through vertical aggradation; (4) travertine deposition occurred in different environments: principally low-energy flat or shallow environments at Tivoli and high-energy inclined environments at Denizli; (5) the growth of the Tivoli plateau occurred in a subsiding basin, whereas the fissure ridges were not influenced by significant subsidence; (6) C- and O-isotope signatures from the two studied travertines are different; (7) despite similar annual precipitations, the present water discharge in the Tivoli area is about ten times greater than that of the Denizli Basin. U-series ages from the two deposits are correlated with paleoclimate oscillations at regional and global scales. Geological field evidence together with paleoclimate correlations suggest that, in both the study cases, the main body of travertine deposits (the bedded travertine) grew preferentially when the water table was high (warm and/or humid periods). Conversely, when the water table was depressed (cold and/or dry periods), the Tivoli travertine underwent partial erosion and the Denizli ridges were cut by axial veins and lateral sill-like structures filled by banded sparitic travertine. A comparative model is proposed where the main factor driving the difference in the morphostratigraphic architecture of fissure ridges and travertine plateaus is the volume of water discharge. A high discharge rate resulted in the precipitation of CaCO3 far away from the springs, hence driving the lateral progradation of the Tivoli plateau. A reduced discharge rate caused travertine precipitation close to the springs, thus causing the vertical aggradation of the Denizli fissure ridges. Paleoclimate oscillations must have controlled the amount of fluid discharge, which, in turn, must have influenced the opening of the feeding fractures by an increased pore pressure.

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1. Introduction

Thermal (or thermogene) travertines are distributed on the Earth's surface with various shapes and sizes, usually in active or recently active tectonic, volcanic, and geothermal areas, where travertines form around geothermal springs. Travertine deposit morphologies as different as cascades, aprons and channels, fissure ridges, plateaus, and towers have been studied in several countries (e.g., Italy, Turkey, and USA; Scholl and Taft, 1964; Buccino et al., 1978; D'Argenio et al., 1983; Hennig et al., 1983; Chaletz and Folk, 1984; Goff and Shevenell, 1987; Altunel and Hancock, 1993a,b; Benson, 1994; Pentecost, 1995; Traganos et al., 1995; Ford and Pedley, 1996; Buchardt et al., 1997; Guo and Riding, 1998, 1999; Atabey, 2002; Özkul et al., 2002; Chaletz and Guidry, 2003; Pentecost, 2005; Crossey et al., 2006; Facenna et al., 2008; Zentmyer et al., 2008; Pedley, 2009; Pedley and Rogerson, 2010; Fouke, 2011; De Filippis et al., in press). These travertine deposit morphologies are in part and in places controlled by the topography over which they formed. Other factors, such as tectonics and climate oscillations, are known or hypothesized to influence the travertine deposition (Bargar, 1978; Sturchio et al., 1994; Çakır, 1999; Hancock et al., 1999; Rihs et al., 2000; Brogi, 2004; Facenna et al., 2008; Mesci et al., 2008; Zentmyer et al., 2008; Selim and Yanik, 2009; Brogi et al., 2010, 2012; De Filippis et al., 2012; Kampman et al., 2012; Nishikawa et al., 2012, 2013) but, in detail, why and how these and other influencing factors control the travertine growth and morphology is still uncertain. It is unclear, in particular, what are the main factors controlling travertine deposits morphologically and volumetrically as different as, for instance, travertine plateaus and fissure ridges (Fig. 1), which have been the object of several studies in the past (Bargar, 1978; Altunel and Hancock, 1993a,b; 1996; Hancock et al., 1999; Rihs et al., 2000; Uysal et al., 2007; Facenna et al., 2008; Brogi and Capezzuoli, 2009) and are again being studied for their numerous implications (De Filippis and Billi, 2012; De Filippis et al., 2012; Gratier et al., 2012; Kampman et al., 2012; Nishikawa et al., 2012; Wigley et al., 2012). Understanding thermal travertine deposition and its influencing factors is, in fact, important, among other reasons, to comprehend the long-term (c. 10^4 a; Uysal et al., 2007; Facenna et al., 2008) geothermal fluid circulation and escape of CO_2 from natural subsurface geologic repositories (e.g., Shipton et al., 2004; Haszdine et al., 2005; Shipton et al., 2005; Anderson and Fairley, 2008; Crosseley et al., 2009; Nelson et al., 2009; Dockrill and Shipton, 2010; Amlan et al., 2011; Gillfillan et al., 2011; Gudmundsson, 2011b; Uysal et al., 2011b; Kampman et al., 2012; Nishikawa et al., 2012, 2013; Wigley et al., 2012). In the last years, several nations have started the practice or at least the planning of subsurface gas storage, including CO_2, either for environmental or for industrial reasons (e.g., Chadwick et al., 2010; Teatini et al., 2011). One of the main threats to such subsurface artificial reservoirs are faulting and fracturing and the related gas seep toward the surface both in a short-term perspective and in a long-term one (Verdon et al., 2011). For this reason, finding and studying geological analogs or markers of geofluid upward circulation are fundamental to determine how and in how much time these fluids can flow through the host rock up to the Earth's surface (Nelson et al., 2009; Uysal et al., 2009; Dockrill and Shipton, 2010).

In this paper, we review and contrast travertine plateaus and fissure ridge travertines (Fig. 1) to understand the reasons for their different growths and morphologies. The term fissure ridge travertines indicates an elongate mound-shaped deposits of thermal travertines, straight or curved in plan view, between a few tens and over 2000 m in length, with a main crestal fissure (Bargar, 1978; Chaletz and Folk, 1984; Altunel and Hancock, 1993a,b; 1996; Çakır, 1999; Hancock et al., 1999; Rihs et al., 2000; Uysal et al., 2007; Facenna et al., 2008; Brogi and Capezzuoli, 2009) and are again being studied for their numerous implications (De Filippis and Billi, 2012; De Filippis et al., 2012; Gratier et al., 2012; Kampman et al., 2012; Nishikawa et al., 2012; Wigley et al., 2012). Understanding thermal travertine deposition and its influencing factors is, in fact, important, among other reasons, to comprehend the long-term (c. 10^4 a; Uysal et al., 2007; Facenna et al., 2008) geothermal fluid circulation and escape of CO_2 from natural subsurface geologic repositories (e.g., Shipton et al., 2004; Haszdine et al., 2005; Shipton et al., 2005; Anderson and Fairley, 2008; Crosseley et al., 2009; Nelson et al., 2009; Dockrill and Shipton, 2010; Amlan et al., 2011; Gillfillan et al., 2011; Gudmundsson, 2011b; Uysal et al., 2011b; Kampman et al., 2012; Nishikawa et al., 2012; Wigley et al., 2012). In the last years, several nations have started the practice or at least the planning of subsurface gas storage, including CO_2, either for environmental or for industrial reasons (e.g., Chadwick et al., 2010; Teatini et al., 2011). One of the main threats to such subsurface artificial reservoirs are faulting and fracturing and the related gas seep toward the surface both in a short-term perspective and in a long-term one (Verdon et al., 2011). For this reason, finding and studying geological analogs or markers of geofluid upward circulation are fundamental to determine how and in how much time these fluids can flow through the host rock up to the Earth's surface (Nelson et al., 2009; Uysal et al., 2009; Dockrill and Shipton, 2010).

![Conceptual sketches for two different morphotypes of travertine deposits: fissure ridges and plateaus. (a) Fissure ridge travertines. This sketch is inspired by a swarm of fissure ridges observed in the geothermal area of Bridgeport (California, USA; De Filippis and Billi, 2012). (b) Travertine plateau. This sketch is inspired by the plateau travertine of the Acque Albule basin (Tivoli, central Italy; Facenna et al., 2008), which is analyzed in this paper.](image)
banded travertine, which is a sparitic, nonporous, often subvertical deposit (i.e., with growth bands of different colors) filling veins injected within the interior part of the bedded travertine or sometimes forming sill-like structures along pre-existing travertine strata (Fig. 1a; e.g., Gratier et al., 2012). Fissure ridge travertines are characterized by a prominent topography as high as circa 30 m (De Filippis et al., 2012). With the term travertine plateau, instead, we refer to a large and massive thermal travertine deposit consisting of bedded travertine filling a tectonic depression and producing no prominent topography (Fig. 1b; Faccenna et al., 2008). We provide further details on these two travertine morphotypes in the following sections. In particular, two key areas are studied (Fig. 2): (1) the travertine plateau of Tivoli in central Italy (Maxia, 1950; Manfra et al., 1976; Chafetz and Folk, 1984; Pentecost and Tortora, 1989; Faccenna et al., 1994c; Pentecost, 1995; Gasparini et al., 2002; Minissale et al., 2002; Pentecost, 2005; Faccenna et al., 2008, 2010; Petitta et al., 2011; Carucci et al., 2012) and (2) the fissure ridge travertines of Denizli Basin in western Turkey (Altunel and Hancock, 1993a,b, 1996; Çakır, 1999; Hancock et al., 1999; Özkul et al., 2002; Altunel and Karabacak, 2005; Uysal et al., 2007, 2009; Kele et al., 2011; De Filippis et al., 2012). We selected these two areas as they are among the most spectacular and well-known endmembers of travertine morphotypes with outstanding exposures and numerous previous studies. The Denizli Basin (Fig. 2a) is located in one of the world’s most rapidly extending regions (e.g., Jackson and McKenzie, 1988; Westaway, 1990), being characterized by an extensional deformation rate of at least 20 mm/a (Reilinger et al., 2006). Present regional heat flow in the area of the Denizli Basin is c. 107 mW/m2 (Mutlu et al., 2008). The extensional tectonic regime of western Turkey is regarded by Sengör et al. (1985) as having started in Tortonian time (c. 7 Ma), when the Anatolian block began to escape toward the west and a N–S-oriented stretching regime started in western Turkey. The principal active normal faults of western Turkey generally strike E–W, but grabens locally trending NW–SE and NE–SW also occur. The

2. Geological setting

2.1. The Denizli Basin

The studied fissure ridges (Kamara, Çukurbag, Akköy, and also a few outcrops of Kocabaş; Fig. 2a) are located in different localities of the Denizli Basin. This basin is situated in western Turkey, at the confluence of the E–W-trending Menderes and NW–SE-trending Gediz grabens, which occur near the eastern margin of the Neogene–Quaternary Aegean extensional province (Westaway, 1990; Bozkurt, 2001) (Fig. 2a). This area is one of the world’s most rapidly extending regions (e.g., Jackson and McKenzie, 1988; Westaway, 1990), being characterized by an extensional deformation rate of at least 20 mm/a (Reilinger et al., 2006). Present regional heat flow in the area of the Denizli Basin is c. 107 mW/m2 (Mutlu et al., 2008). The extensional tectonic regime of western Turkey is regarded by Sengör et al. (1985) as having started in Tortonian time (c. 7 Ma), when the Anatolian block began to escape toward the west and a N–S-oriented stretching regime started in western Turkey. The principal active normal faults of western Turkey generally strike E–W, but grabens locally trending NW–SE and NE–SW also occur. The

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![Fig. 2. Study areas. (a) Simplified geological map of the Denizli Basin, western Turkey. Red stars indicate the location of the studied fissure ridges. (b) Simplified geological map of the Acque Albule basin, Tivoli, central Italy, showing the studied travertine plateau. See also Fig. 1(b). (For interpretation of the references to color in this figure, the reader is referred to the web version of this article.)](image-url)
Neogene Denizli Basin trends NW–SE with a length of about 50 km and a width of 20 km, and rests at an altitude of about 200 m a.s.l. The metamorphic rocks exposed in the Menderes Massif constitute the basement of the Denizli Basin, whose formation started in early Miocene time (Ağıcık et al., 2007). Marbles are an important constituent of the metamorphic basement (Pamir and Erentöz, 1974; Okay, 1989) and are, together with Pliocene continental limestone, at the origin of the studied travertines (Uysal et al., 2007). An important locality in the Denizli Basin is Pamukkale (Turkish for Cotton Castle), where actively accumulating travertines are one of Turkey’s most famous tourist sights (Fig. 2a). Travertine deposition at Pamukkale has been active since at least 400 ka, partially covering the Roman city and necropolis of Hierapolis (Altunel and Hancock, 1993b). At present, in the Pamukkale area, the thermal water emerges at temperatures between 13.3 and 51.3 °C from springs along fractures (Kele et al., 2011), whereas, in general in the Denizli Basin, water temperatures are between 28 and 59 °C (Gökgoz, 1998; Özlter, 2000; Dilsiz et al., 2004). In the Pamukkale area, about 17 warm springs discharge waters of the Ca–Mg–HCO3–SO4-type with electrical conductivity of c. 3.0 mS/cm (Gökgoz, 1998) and a total average discharge between 365 and 385 l/s (Dilsiz, 2006). In the same area (Pamukkale), the average annual precipitation is around 837 mm/a (Dilsiz, 2006).

### 2.2. The Acque Albule basin

The Acque Albule basin (Tivoli, central Italy), famous in the world for the Lapis Tiburtinus travertine, is located to the north of the quiescent Colli Albani volcano (Funicelli et al., 2003; Facenna et al., 2008, 2010), along the western boundary of the Neogene central Apennines fold–thrust belt (Fig. 2b). In this side of the Apennines belt, reduced thickness of the lithosphere, volcanism, extensional basins, and high heat flow (average values between 100 and 200 mW/m2; Mongelli and Zito, 1991) are the results of the Neogene–Quaternary backarc and post-orogenic extensional processes (Funicelli et al., 1976; Chiodini et al., 2004; Acocella and Funicelli, 2006). At present, the Acque Albule basin is mainly characterized by N-striking right-lateral and NE-striking transtensional-to-normal faults (Alfonsi et al., 1991; Facenna et al., 1994a). These structures have partially controlled the latest stages of volcanism and related hydrothermal outflows (e.g., Billi et al., 2007). These same structures generated the Pleistocene pull-apart depression of the Acque Albule basin, which has been filled by the Lapis Tiburtinus travertine resulting among the largest and thickest travertine plateaus in the world (Fig. 2b: Facenna et al., 1994b, 2008). The source of the CaCO3 forming the travertine deposit is principally the subsurface Meso-Cenozoic carbonate succession (Manfra et al., 1976; Minissale et al., 2002), which is widely exposed to the north and east of the Acque Albule basin. The presence of the Meso-Cenozoic carbonate succession below the nearby Colli Albani volcano is also known from geophysical and volcanic evidences (e.g., Funicelli et al., 1976). Travertine deposition in the Acque Albule basin started during the middle–late Pleistocene time along the N- and NE-striking right-lateral strike-slip and transtensional faults (Fig. 2b). Some of these faults are still active (or simply activated by pore pressure increases) as demonstrated by the low-magnitude, shallow (less than 1.5 km), seismic sequence that occurred in 2001 beneath the Acque Albule basin (i.e., right beneath the travertine plateau; Gasparini et al., 2002; see also F republici et al., 2010). Moreover, two samples of spuritic calcite filling some extensional fractures along the above-mentioned faults were dated to circa 49 and 59 ka, which is a time contemporaneous with the plateau travertine deposition (Faccenna et al., 2008).

The Acque Albule springs, which are located on top of the travertine plateau (Fig. 2b), have a mean discharge ranging between 2 and 3 m3 s–1 (Capelli et al., 1987; Facenna et al., 1994c) and form two deep lakes (Regina and Colonnelle lakes) with waters at temperatures of c. 23–24 °C (Petitta et al., 2011). Other minor springs in the area together with the Aniene River streambed springs all contribute with an additional 1 m3 s–1 to the water discharge in the area (Petitta et al., 2011), where the average annual precipitation is around 832 mm/a (Capelli et al., 1987). In the Regina and Colonnelle lakes, Petitta et al. (2011) measured the following physicochemical parameters for the thermal waters: T = 23.5 °C, pH = 6.2, EC = 2.63 mS cm–1, Eθ = –294 mV, pCO2 = 915 atm × 10–3, whereas Capelli et al. (1987) measured a salinity of 2670 and 2450 mg/l, respectively, for the waters of the Regina and Colonnelle lakes, with a temperature of 23 °C for both lakes.

### 3. Methods and results

#### 3.1. Fissure ridge travertines (Denizli Basin, Turkey)

Field observations and U–Th geochronological data concerning three studied fissure ridge travertines (Kamara, Çukurbag, and Akköy) from the Denizli Basin are synthesized below after De Filippis et al. (2012) and previous papers (Altunel and Hancock, 1993a,b; Çakir, 1999; Hancock et al., 1999, 2002; Altunel and Karabacak, 2005; Uysal et al., 2007, 2009). A common characteristic of all studied fissure ridges in the Denizli Basin is that we did not observe faults through them; rather, we observed axial and lateral fissures either open or filled by banded travertine. Volume of fissure ridge travertines varies with their size up to a maximum of about 10–3 km3. From a sedimentary point of view, the banded travertine from the studied fissure ridges grew as flowstone on sloping surfaces forming lithofacies such as shrubs, crystalline crusts, and lithoclasts. These facies are typical of high energy slope or even cascade environments, where travertine grew with a shallow-to-steep clinostратification. This primary inclination of bedding is partly preserved in the flanks (partly tilted after deposition or during its late stages; De Filippis et al., 2012) of the studied fissure ridges. Lithofacies typical of low energy environments such as large pools are rarer (Atabay, 2002; Özkul et al., 2002). Most fissure ridges are characterized by a vertical aggradational pattern of strata (bedded travertines). The most striking example of aggradational travertine strata that we could observe is exposed in the Kamara fissure ridge (Figs. 3a and 4a). As an additional instance, in Fig. 5, we also report an outstanding example of aggradational travertine strata in the BRP1 fissure ridge from the Bridgeport (California) geothermal area (De Filippis and Billi, 2012). In the first example (Kamara in Figs. 3a and 4a), an upward thickening of strata along the crest of the fissure ridge is evident. The upward aggradation is accompanied by the progressive upward steepening of fissure ridge flanks reflected in progressively-higher-energy depositional facies (i.e., in Fig. 3a, see the fossil waterfall in the uppermost part of the northeastern flank). The second example (BRP1 in Fig. 5) is characterized by a fan-shaped pattern of bedded travertine along the southwestern flank, with obvious upward thickening and steepening of strata. Concerning the shape of fissure ridges and their aggradational pattern, it is useful to note that De Filippis and Billi (2012), considering 33 fissure ridges from various localities in the world (including Denizli), found that the height-to-width aspect ratio of fissure ridges is normally around 0.2.

The Kamara fissure ridge, part of the Yenice travertine deposit, is located on the Tripolis fault footwall (Çakir, 1999), which is, in turn, the hangingwall of other normal faults located northeast of Kamara (Yalınlar, 1983; Kaymakçı, 2006) (Fig. 2a). Kamara is an active fissure ridge, where travertine is being deposited from emerging carbonate-rich hot waters. The fissure ridge (Figs. 3a, 4a, and b) is N125°-trending (the long axis), c. 63 m in length, 15 m in width, and 6 m in maximum height over the surrounding plain. Fig. 3(a) displays a cross-section through the fissure ridge, showing an asymmetric profile with the northeastern flank slightly steeper than the southwestern one. At the top of the ridge, along its crest, we observed a system of en-echelon axial open fractures (Fig. 4a) with an average...
strike of about N120°. Both flanks of the fissure ridge consist of bedded travertines dipping away from the axial fissure. The Kamara fissure ridge is strongly asymmetrical due to the presence of a fossil waterfall along the northeastern flank (Figs. 3a and 4a). The Kamara banded travertine is exposed along the southwestern side of the axial fissure (Fig. 4b). The contact between bedded and banded travertines, when visible, is nearly orthogonal, with the banded travertine cutting through (i.e., injected within) the bedded one (Fig. 4b). At least in one case, we observed a set of banded travertine layers curving to form a sill-like structure intruded along the strata of the bedded travertine (Fig. 4b; e.g., Gratier et al., 2012). A further spectacular example of sill-like structure along travertine bedding is exposed in the Kocabas fissure ridge (Fig. 4f and g), which is located in the southeastern sector of the Denizli Basin (Fig. 2a; compare with similar volcanic structures in Gudmundsson, 2011a). U-Th dating analysis provided ages of c. 1.7 and 2.5 ka (De Filippis et al., 2012) for two samples of banded travertine from the Kamara fissure ridge (Fig. 5). The adjacent bedded travertine is too detrital to be reliably dated. The temperature of the Kamara spring has decreased from 57 to 32 °C in the period 1976–1998. In this area, in 1998, a well was drilled to a depth of 145 m. The well initially discharged 4 l/s and banded travertine is visible only in the northernmost part of the Pamukkale travertine deposit (Fig. 2a). This part of the Denizli Basin is characterized by the following physicochemical parameters: T = 56.1 °C, pH = 6.9, EC = 3.01 mS cm⁻¹, Eh = 4.3 mV, free CO₂ content = 960 ppm, and alkalinity = 1178 (Kele et al., 2011).

The Akköy fissure ridge (Karakaya Hill) constitutes the northern portion of the Pamukkale travertine deposit (Fig. 2a). This part of the Denizli Basin is characterized by a set of fissure ridges, either active or extinct, mostly grown in the step-over zone between the Akköy and Hierapolis faults, where an intense fracture network likely
enhanced the leakage of mineralizing fluids. The Akköy fissure ridge is inactive and located on the hangingwall of the Hierapolis Fault (Fig. 2a). Below, we present observations made on the northern segment of the Akköy fissure ridge (De Filippis et al., 2012), where excellent exposures of travertine are available in active quarries (Fig. 4d and e).

The Akköy fissure ridge trends c. N145° with a length of about 1900 m, a width of 200 m, and an average height of 25 m. The Akköy structure is one of the biggest fissure ridge hitherto known on the Earth (De Filippis et al., 2012). Fig. 3(c) shows a transverse cross-section of the Akköy ridge. Both main and minor fractures are nearly vertical in the central portion of the fissure ridge, and become inclined toward the northeast, forming a fan-shaped pattern (Fig. 3c and d). These fractures are filled by banded travertine (at present almost entirely quarried; Fig. 4d) and are flanked by bedded travertine, which dips away from the axial zone. In general, the fissures filled by banded travertine tend to widen upward and become inclined toward the northeast from a subvertical attitude in the central portion of the ridge. One important feature observed in the cross-section of Fig. 3(c) is a bedded travertine unconformity marking the contact between two adjacent fissure ridges coalesced to form this branch of the Akköy fissure ridge. Except for this unconformity, in the Akköy fissure ridge as well as in the other studied ones, we did not observe evidence of manifest erosional surfaces accompanied by paleosoil and kart features, which are, in contrast, frequent in the Tivoli travertine plateau. U–Th geochronological analyses (De Filippis et al., 2012) provided ages between c. 16.3 and 29.3 ka for the banded travertine, and ages of 36.4 and 53.4 ka for two bedded travertine samples collected along the cross-section of Fig. 3(c). Uysal et al. (2007, 2009) dated the banded travertine within the southeastern branch of Akköy fissure ridge (south of our study area) and obtained ages between c. 48.7 and 73.6, whereas ages obtained by the same authors on the banded travertine from our study area are between c. 21.0 and 25.6 ka. Moreover, Altunel and Karabacak (2005) obtained an age of 34.9 ka for a single sample from the banded travertine in our study area.

Age data from the fissure ridges studied in this paper and in previous ones indicate that the deposition rate of the bedded travertine is usually around 0.2–0.3 mm/a (Altunel and Hancock, 1993b), whereas the rate for the banded travertine is usually ten times smaller (i.e., circa 0.03 mm/a; Uysal et al., 2007; Mesci et al., 2008). This latter value is consistent with the order of magnitude for post-mortem average opening rates of six fissure ridges from the Denizli Basin (Altunel and Karabacak, 2005). Age data also indicate that the longevity of fissure ridges is usually on the order of 104 years (Altunel and Karabacak, 2005; Uysal et al., 2007, 2009).

3.2. Travertine plateau (Tivoli, Italy)

Field observations, subsurface data, and U–Th geochronological results concerning the travertine plateau of Tivoli (Acque Albule basin)
are synthesized below after Faccenna et al. (2008, 2010) and previous papers (Maxia, 1950; Chafetz and Folk, 1984; Pentecost and Tortora, 1989; Faccenna et al., 1994c).

The travertine plateau of Tivoli (Fig. 6) fills a transtensional tectonic depression (the Acque Albule basin) generated by the Quaternary activity of N-S right-lateral faults and associated NE–SW transtensional and normal ones (Faccenna et al., 1994c). The upper surface of the travertine plateau is substantially flat or shallow, thus producing no prominent topography (Fig. 7b). The travertine plateau is characterized by marked lateral variations in thickness: from 40 to 50 m on average in the central sector up to only a few meters or few tens of meters toward the east, the north, and the south, where the travertine deposit laterally closes (Fig. 6). Toward the west, the travertine thickens up to a maximum of about 90 m. This travertine thickening corresponds to a N-striking active fault below the travertine plateau (Faccenna et al., 1994c; Frepoli et al., 2010) and, on top of the travertine deposit, to the N-S alignment of thermal springs, gas vents, and sinkholes (Maxia, 1950) (Fig. 6). In the map view, the travertine plateau is about 7 by 5 km for a total volume on the order of 1 km$^3$.

The Tivoli travertine (Fig. 7) includes finely laminated carbonates with alternations of bacterial mud tufa-like travertines and laminated facies (Chafetz and Folk, 1984; Pentecost and Tortora, 1989). These sedimentological features are indicative of travertine growth in very-low-energy-environments such as restricted pools or lakes a few meters deep at most (i.e., usually less than about 1–2 m). In these environments, the travertine strata usually grow with a sub-horizontal attitude.

At Tivoli, high-energy travertine facies are present only in some places. The Tivoli travertine deposit is composed of a sequence of sub-horizontal to gently southward dipping depositional units or benches, which are c. 7–10 m in thickness (Fig. 7). These benches are separated by gently-southward dipping erosional surfaces usually marked by thin (less than about 1 m) soil deposits associated with several karst features (e.g., sinkholes and other karst erosional features) and pockets of conglomerates (Fig. 7a to c). The erosional surfaces are, in places, subparallel to the depositional benches, but, more often, are oblique, cutting through the benches and also through older erosional surfaces. So far, no in situ banded travertine has been observed in the Tivoli area. The bench sedimentology and attitude are indicative of travertine strata grown in a sub-horizontal to gently-southward-dipping attitude with the bottom of each travertine bench onlapping the underlying erosional surface and bench (Figs. 6, 7a, and b). This evidence indicates the occurrence, after each erosional episode, of a new cycle of travertine deposition in sub-horizontal to pools and terraces occurring over the inclined erosional surface. Moreover, the occurrence of erosional surfaces with associated karst features and pockets of clay and conglomerate is the manifest symptom of a lowering of the water table in the depositional basin during erosional periods. On the other hand, the water table must have been very close to the depositional surface during depositional periods (Faccenna et al., 2008).

The Tivoli travertine is particularly compact and lithoid such that this rock has been largely used as a constructive and decorative stone since at least the Roman age (e.g., the Colosseum in Rome is largely made of Tivoli travertine). An exception is the so named testina layer (c. 3–4 m in average thickness) consisting of a poorly compact and poorly lithoid travertine capping most part of the Tivoli plateau (Fig. 8a). The testina layer is the youngest travertine of the Tivoli plateau (29 ± 4 ka, Faccenna et al., 2008).

Several quarry exposures together with borehole data made available from the local quarry industry allowed Faccenna et al. (2008) to reconstruct the geometry of the travertine deposit and erosional surfaces in three-dimension. The travertine deposit has a typical progradational pattern with a gentle upward steepening of strata toward the south. This evidence indicates a travertine growth farther out toward the Aniene River (south) over time (see Fig. 7b and the A–A’ cross-section in Fig. 6). The height-to-width aspect ratio of the Tivoli plateau is circa 10$^{-2}$, which is one order of magnitude smaller than the one typical of fissure ridges (De Filippis and Billi, 2012). Five main erosional surfaces, termed S1, S2, S3, S4, and S5 from the youngest surface to the oldest one, are reconstructed (Figs. 6, 7a, and b). The age of erosional surfaces is constrained by the ages of travertine samples gathered immediately below and above the erosional surfaces (Faccenna et al., 2008). Results indicate that S1 dates to 34 ± 5 ka, S2 to 44 ± 4 ka, S3 to 56.5 ± 8 ka, S4 to 82 ± 9 ka, and S5 to 99 ± 5 ka. The longevity of the Tivoli travertine is c. 100 ka at least, which is slightly longer than that typical of several fissure ridges (on the order of 10$^4$ years). Considering the average thickness of the Tivoli plateau, we obtain an approximate average deposition rate of 0.4 mm/a, which should approximately correspond with the average vertical subsidence rate of the Acque Albule basin or with an overvalue of this subsidence (the travertine progradation pattern, in fact, suggests a moderate basin subsidence). Assuming this subsidence rate (0.4 mm/a) as due to motion along basin-bounding normal faults and considering these normal faults as dipping by 60°, we obtain a horizontal extension (heave) rate of about 0.25 mm/a (= 40,000 mm × tan 30°/100,000 years). This value should, however, be reduced by a percentage due to the portion of subsidence induced by sediment compaction.

3.3. Isotope data (Tivoli and Denizli Basin)

Stable isotope analyses have proved to be relevant to understand the genesis of travertines (Friedman, 1970; Manfra et al., 1976; Turi,
Carbon and oxygen isotopic compositions can provide insights into the precipitation conditions, the origin of carbon, and the provenance of the travertine source fluids. To define the isotopic composition of the bedded travertine from Tivoli, we collected 31 samples from an E–W section (c. 1 km in length) through the southern part of the travertine plateau. Samples, in particular, were collected above and beneath the S3 surface (c. 56 ka; Figs. 6 and 7). Sample locations are schematically shown in Fig. 6. In the Denizli Basin, we collected and analyzed a total of 15 samples for the banded travertine and 10 samples for the bedded travertine from the studied fissure ridges (De Filippis et al., 2012). All results are reported in Fig. 8 and Table S1. Isotope data for the Tivoli travertine are more concentrated in a small range, whereas data for the Denizli travertines are slightly more dispersed (Fig. 8). In detail, at Tivoli, delta units are between $-7.18$‰ and $-4.76$‰ for oxygen and $+8.31$‰ and $+10.77$‰ for carbon. At Denizli, delta units are between $-14.1$‰ and $-7.1$‰ for oxygen and $+2.9$‰ and $+7.8$‰ for carbon.

In a previous study, Manfra et al. (1976) analyzed oxygen and carbon isotopes of travertine samples from Tivoli and obtained results similar to those reported in Fig. 11. Manfra et al. (1976) attributed the limited dispersion of the oxygen values to isotopic resetting under diagenetic dissolution and reprecipitation processes at temperatures near 15 °C. This temperature was calculated using the paleotemperature equation of O’Neil et al. (1969) and assuming an isotopic composition of the precipitation water in the range of the local meteoric water ($-6 \div -8$‰ vs. SMOW). Diagenetic processes
of compaction and cementation of the Tivoli travertine are evident also from the marked difference between the recent testina layer (poorly compact) and the older travertine (very compact) underlying the testina layer (Fig. 7a). Manfra et al. (1976) highlighted an isotopic difference in oxygen isotopes between the testina layer and the underlying travertine. Delta units for the testina layer are between $-4.5$ and $-4.0 \permil$, whereas delta units for the underlying travertine are between $-7.18 \permil$ and $-4.76 \permil$. On the basis of this difference, it is reasonable to contend that the testina layer may isotopically represent a pristine poorly-diagenized travertine. In contrast, the more compact and lithoid underlying travertine is at least in part the product of posterior diagenetic alteration. We argue that the diagenesis of the Tivoli travertine (except the testina layer) has obliterated the original oxygen isotopic signature and, consequently, the isotope variability. Such variability was probably caused by temperature fluctuations during deposition and by precipitation waters of a different origin. The high-positive values for carbon isotopes of the Tivoli travertine (Fig. 8) indicate that the precipitation waters were rich in CO$_2$ derived from decarbonation of marine carbonates, which are largely present in the Tivoli substratum. Also Minissale et al. (2002), on the basis of chemical and isotopic studies on the Tivoli travertine, found an origin by spring waters heated during transit in a high heat-flow area and enriched by a large quantity of CO$_2$ derived from decarbonation of limestones in the substratum. The warm deep waters must have intercepted, during their upward rising, the colder shallow aquifer mainly recharged by meteoric precipitation. The meteoric cold waters must have lowered the temperature of the rising deep waters and reset the oxygen isotopes of the travertine during diagenetic processes.

Concerning the travertine samples from the Denizli Basin, the isotope data were discussed in detail in De Filippis et al.’s (2012) study. The variability of oxygen isotope composition is the expression of a precipitation from aqueous solutions generated by the mixing, at a variable extent, between the meteoric waters (low isotope composition) and the deep geothermal fluids (high isotope composition). The interaction between these two fluid reservoirs is described in detail by Dilisz et al. (2004) and Dilisz (2006). The range in carbon isotopes of the Denizli travertines shows that the precipitation waters had similar characteristics of those waters that fed the Tivoli travertine. Both waters, in particular, were rich in CO$_2$ derived from decarbonation of the underlying carbonate rock.

3.4. Correlation of U-series ages with global and regional climate events

Before approaching the correlation of our U-series ages with global and regional climate events, we acknowledge, as it is well known from several previous studies, that travertine dating can be affected by large error bars connected with the presence of detrital Th.
in the samples, particularly in the poorly sparitic bedded travertine (e.g., Facenna et al., 2008; Tuccimei et al., 2010). We acknowledge also that the above-presented field evidence (Figs. 3, 4, 6, and 7) shows that both the Tivoli plateau and the Denizli fissure ridges grow with a cyclical pattern consisting of (1) deposition vs. erosion phases, in the Tivoli case, and (2) bedded vs. banded travertines deposition phases, in the Denizli case. In both cases, these cycles are symptomatic of vertical fluctuations of the water table in the deposition basins. In particular, in the Tivoli case, the erosional surfaces across the travertine plateau are always accompanied by frequent karst features (i.e., particularly below the erosional surfaces; Fig. 8). These features, together with the erosional surfaces, are unequivocally symptomatic of a significant lowering of the water table in the deposition basin during the erosional events (Facenna et al., 2008). In the Denizli case, the growth of the banded travertine within the studied fissure ridges are symptomatic of a poor water discharge and, therefore, of a low-stand of the water table as also constrained by Uysal et al. (2009). It follows that, regardless of the reliability of the correlation between our U-series ages and global and regional climate events (depending on the accuracy of U-series ages), our field evidence shows a clear correlation between the cyclical growth of the studied travertines and water table fluctuations. These fluctuations, in turn, should have been conceivably controlled by paleoclimate oscillations. Future improvements to travertine dating methods will help to enhance this latter notion, which is addressed as follows.

To understand the influence of paleoclimate on the growth of Denizli and Tivoli travertines, in Figs. 9 and 10, we compare the available radiometric age data (Table S2) with major paleoclimate indicators and events determined both at the global and regional scales during Quaternary time (Tzedakis et al., 1997, 2001; Wang et al., 2001; Tzedakis et al., 2006; Wanner et al., 2008; Fleitmann et al., 2009; Fletcher et al., 2010; Stenni et al., 2010; Büntgen et al., 2011; Uysal et al., 2012; Clark et al., 2012; see also Uysal et al., 2007; Facenna et al., 2008; Uysal et al., 2009; De Filippis et al., 2012). In particular, as paleoclimate indicators, we use records extracted from Greenland ice cores and speleothems from China (Hulu cave) and Israel (Sanbao cave) (GISP2, 1997; Bar-Matthews et al., 1999; Wang et al., 2001, 2008). Data from a stalagmite of Sofular Cave (northern Turkey), which is rather close to the Denizli Basin, are in excellent agreement with the Hulu Cave data (Fleitmann et al., 2009; Badertscher et al., 2011). Temperature changes during the last 90 ka are determined from high-resolution analyses on Antarctica ice cores (Fig. 9a; Stenni et al., 2010; Badertscher et al., 2011). Temperature changes during the last 90 ka are determined from high-resolution analyses on Antarctica ice cores (Fig. 9a; Stenni et al., 2010; Badertscher et al., 2011). For the last 2500 years (Fig. 9c), we use the oscillations of precipitation totals and temperature anomalies in Europe determined by Büntgen et al. (2011). For a possible correlation with precipitation (humid vs. dry climate), we use pollen records (Fig. 10) from Valle di Castiglione (only a few kilometers to the south of Tivoli) and from Ioannina and Tenaghi Philippon (eastern Mediterranean) (Tzedakis et al., 2001, 2006; Fletcher et al., 2010). Pollen curves represent the ratio (i.e., in percentage) between arboreal and non-arboreal pollens (Fig. 10). It follows that peaks can be interpreted as corresponding to relatively humid times (more arboreal pollens), whereas depressions represent relatively dry times (less arboreal pollens). The ages of most banded travertine samples from Denizli fissure ridges coincide with cold (dry) events. In particular, Fig. 9(b) shows that the U-series dates for the banded travertine fall within the 16–30 ka time interval, with 6 out of 8 samples coinciding with or being very close to cold climate events such as LGM, H1, and H2 (Bar-Matthews et al., 1999; Wang et al., 2001). Other samples fall within or very close to older cold events such as H5, H6, and others (i.e., between 14 and 15 and between 20 and 21 along the Sanbao cave curve, Fig. 9b). The age of the two youngest samples (Kamara, 1.7 ± 0.1 and 2.5 ± 0.1 ka; Table S2) corresponds, in the youngest case (i.e., 1.7 ka), with a well-known dry event (Fig. 9c; Orland et al., 2011; Büntgen et al., 2011; Orland et al., 2012) and, more in general, with a dry period as also demonstrated by the contemporaneous strong depression of the Lake Lisan level (i.e., the level of the paleo-Dead Sea; Bartov et al., 2003). The general temporal correlation between banded travertine deposition in the Denizli Basin and cold climate events was already ascertained by Uysal et al. (2007, 2009), whose geochronological data are also plotted in Figs. 9 and 10. Two age data around 28 ka (Fig. 9b) do not correlate with the major cold events, but the associated error bars are too large to prove or disprove such a general correlation.

Fig. 10 shows that the age of most banded travertine samples fall in dry periods younger than 80 ka. Less evident is the correlation concerning samples with ages between about 80 and 120 ka, which is a period generally humid but with several dry events. On the contrary, two age data between about 150 and 160 ka fall in dry periods as demonstrated by the Tenaghi Philippon curve (Fig. 10). Concerning the bedded travertine (from both Denizli and Tivoli), as abovementioned, the main problem with the geochronological data is that they are affected by rather large error bars (Table S2) connected with the presence of detrital material within the samples. For this reason, the following correlations are indicative and necessitate further corroboration in the future. Concerning radiometric ages of the bedded travertine from Denizli (Table S2), samples Ak27 and Ak24 provided ages of 36.4 ± 3.0 and 53.4 ± 1.6 ka, respectively. These ages (c. 36 and 54 ka) correspond with warm peaks during Quaternary time (Bar-Matthews et al., 1999). A third sample (Pa5) provided an age of 60 ± 1 ka, which corresponds with the onset of
Figure 9
the warm (and relatively wet) MIS 3 (Martinson et al., 1987).

Samples of bedded travertine from the Tivoli plateau were collected close to the erosional surfaces (where possible, both above and below them) to constrain the ages of the erosional surfaces themselves (named S1, S2, S3, S4, and S5, Figs. 9 and 10; Faccenna et al., 2008). Correlations with the Valle di Castiglione pollen curve (Fig. 10) suggest that at least three erosional surfaces (S1, S2, and S3) developed under relatively dry events. Also the S4 surface may have been developed during a dry period, but the large error bar associated with this datum makes this correlation uncertain. In contrast, the age of the S5 surface seems to fall in a relatively humid time.

4. Discussion and conclusions

The overall genesis and growth of the Denizli and Tivoli Quaternary travertines are known from several previous works (Altunel and Hancock, 1996; Uysal et al., 2007; Faccenna et al., 2008; Uysal et al., 2009; Faccenna et al., 2010; De Filippis et al., 2012). With the parallel study presented in this paper, we rather focus on the differences and similarities between the Tivoli plateau and Denizli fissure ridges in the attempt to understand the reasons by which a geothermal circuit can lead to the deposition of travertine morphotypes as different as those here investigated (Fig. 1). To address this problem, in Table 1, we have synthesized all main geological features concerning the studied travertines and their boundary conditions. Despite similar (i) bulk lithologies...

Fig. 10. Figure modified after Tzedakis et al. (2001) and references therein. Comparison of U/Th ages (diamonds with horizontal error bars = 2σ) of the bedded and banded travertines from the Acque Albule (Tivoli) and Denizli basins with the alignment of terrestrial pollen records to the marine isotopic sequence. For the terrestrial records, the arboreal (AP) minus Pinus (solid line) curve has been relied for correlations and tuning, but the AP curve including Pinus (dotted line) is also shown. The curves represent the ratio (i.e., in percentage) between arboreal and non-arboreal pollens. It follows that peaks can be interpreted as corresponding to relatively humid times (more arboreal pollens), whereas depressions represent relatively dry times (less arboreal pollens). Data are from Valle di Castiglione (only about 10 km from the Tivoli travertine) and from Ioannina and Tenaghi Philippion (eastern Mediterranean) (Tzedakis et al., 2001, 2006; Fletcher et al., 2010). Marine isotope stages during late Pleistocene time (Martinson et al., 1987) are also indicated. Glacial times correspond to depressions, whereas interglacial times correspond to peaks.
Plateau versus fissure ridge travertines: comparison between main attributes. Note that data with asterisk refer to the present time and not to the time of travertine formation.

Table 1

<table>
<thead>
<tr>
<th></th>
<th>Travertine plateau (Tivoli)</th>
<th>Fissure ridge travertines (Denizli)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Geometry</strong></td>
<td>Large size</td>
<td>Small size (single ridge)</td>
</tr>
<tr>
<td></td>
<td>Area c. 30 km²</td>
<td>Area c. 1 km²</td>
</tr>
<tr>
<td></td>
<td>Volume c. 1 km³</td>
<td>Volume c. 0.01 km³</td>
</tr>
<tr>
<td><strong>Geomorphology</strong></td>
<td>Flat or shallow</td>
<td>Prominent ridges</td>
</tr>
<tr>
<td></td>
<td>Presence of obvious</td>
<td>No obvious erosional surfaces</td>
</tr>
<tr>
<td></td>
<td>through-going erosional</td>
<td></td>
</tr>
<tr>
<td></td>
<td>surfaces</td>
<td></td>
</tr>
<tr>
<td><strong>Travertine type</strong></td>
<td>Bedded travertine, no</td>
<td>Bedded travertine by veins</td>
</tr>
<tr>
<td></td>
<td>banded travertine</td>
<td></td>
</tr>
<tr>
<td><strong>Deposition environment</strong></td>
<td>Mostly low-energy lake/</td>
<td>Heterogeneous environments, usually</td>
</tr>
<tr>
<td></td>
<td>pool environment</td>
<td>high-energy (cascade, slope)</td>
</tr>
<tr>
<td></td>
<td>Cycle deposition vs. erosion</td>
<td>Cyclic, deposition of bedded vs.</td>
</tr>
<tr>
<td></td>
<td>of bedded travertine; lateral</td>
<td>banded travertine; vertical</td>
</tr>
<tr>
<td></td>
<td>progradation of strata</td>
<td>aggradation of strata</td>
</tr>
<tr>
<td><strong>Source rock</strong></td>
<td>Meso-Cenozoic marine</td>
<td>Marbles from metamorphic basement</td>
</tr>
<tr>
<td></td>
<td>carbonates</td>
<td>and Phocene continental limestone</td>
</tr>
<tr>
<td></td>
<td>$v^{13}C$ (PDB) = c. 8 to 11</td>
<td>$v^{13}C$ (PDB) = c. 3 to 8</td>
</tr>
<tr>
<td></td>
<td>$v^{18}O$ (PDB) = c. 7 to -4</td>
<td>$v^{18}O$ (PDB) = c. -14 to -8</td>
</tr>
<tr>
<td><strong>Tectonics</strong></td>
<td>Strike-slip pull-apart</td>
<td>Extensional (basin) tectonics</td>
</tr>
<tr>
<td></td>
<td>tectonics</td>
<td></td>
</tr>
<tr>
<td></td>
<td>*Weakly active tectonics</td>
<td>*Regional active tectonics</td>
</tr>
<tr>
<td></td>
<td>perhaps solely induced by</td>
<td>(regional extension rate = c. 20 mm/a)</td>
</tr>
<tr>
<td></td>
<td>fluids</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Faults below and around</td>
<td>Fissures (no faults) through ridges</td>
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<tr>
<td></td>
<td>travertine plateau</td>
<td></td>
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<tr>
<td></td>
<td>No injection veins, banded</td>
<td>Injection veins, banded travertine,</td>
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<td></td>
<td>travertine, and sills</td>
<td>and sills</td>
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<tr>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Source water</strong></td>
<td>c. 100/200 mM/m²</td>
<td>c. 107 mW/m²</td>
</tr>
<tr>
<td><strong>Heat flow</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td><em>Discharge</em></td>
<td>2–3 m³/s</td>
<td>0.3–0.4 m³/s</td>
</tr>
<tr>
<td><strong>Temperature</strong></td>
<td>22–24 ºC</td>
<td>28–59 ºC</td>
</tr>
<tr>
<td><strong>Electrical conductivity</strong></td>
<td>c. 2.6 mS cm⁻¹</td>
<td>c. 3.0 mS cm⁻¹</td>
</tr>
<tr>
<td><strong>Feeding conduit</strong></td>
<td>Uncertain, probably along</td>
<td>Evident mineralized axial conduits</td>
</tr>
<tr>
<td></td>
<td>faults below plateau</td>
<td>(banded travertine)</td>
</tr>
<tr>
<td><strong>Average annual precipitation</strong></td>
<td>c. 832 mm/a</td>
<td>c. 837 mm/a</td>
</tr>
<tr>
<td><strong>Climate influence</strong></td>
<td>Bedded travertine</td>
<td>Bedded travertine</td>
</tr>
<tr>
<td></td>
<td>Growth mainly during warm</td>
<td>Growth mainly during cold times</td>
</tr>
<tr>
<td></td>
<td>times</td>
<td></td>
</tr>
<tr>
<td><strong>Maximum longevity</strong></td>
<td>c. 100 ka</td>
<td>c. 50 ka</td>
</tr>
<tr>
<td><strong>Average deposition rate</strong></td>
<td>c. 0.4 mm/a (vertical)</td>
<td>c. 0.3 mm/a (vertical)</td>
</tr>
<tr>
<td></td>
<td>Banded travertine</td>
<td>Banded travertine</td>
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<td></td>
<td>Banded travertine</td>
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</table>

(i.e., the bedded travertine), (ii) longevities (c. 10⁴–10⁵ years; Table 1), (iii) average vertical deposition rates (c. 0.3–0.4 mm/a; Table 1), and (iv) cyclic growth patterns in both cases, we observe the following significant differences, which are relevant to understand the differential evolution of the studied travertines (Table 1):

(1) For a similar longevity and vertical deposition rate, the volume of the Tivoli plateau is about one hundred times larger than each fissure ridge of the Denizli Basin. This implies vastly different volumetric deposition rates between the studied travertine deposits;
(2) Despite similar bulk lithologies (bedded travertine), the travertine plateau does not produce relief or so, whereas the fissure ridges are characterized by their prominent topography consisting of an elongate mound-shaped relief along the feeding fissure. Paradoxically, a prominent topography is characteristic of fissure ridges that are, volumetrically, far smaller than the flat travertine plateaus;
(3) The different volumes and morphologies of fissure ridges and plateaus are a result of different depositional environments and growth styles: principally low-energy flat or shallow environments at Tivoli (pools and lakes), where the travertine plateau grew mainly by lateral progradation, and high-energy inclined environments at Denizli (slopes and cascades), where the fissure ridge travertines grew mainly by vertical aggradation;
(4) The growth of the Tivoli plateau occurred in a subsiding basin controlled by dip-slip fault motion and sediment compaction for a horizontal deformation rate estimated as ≤0.25 mm/a. In contrast, the studied fissure ridges did not grow up on a locally subsiding substratum and the opening rates of their axial vertical fissures are only on the order of 10⁻² mm/a or even less (Altunel and Karabacak, 2005; Uysal et al., 2007; Mesci et al., 2008). Note, however, that some differential limited subsidence occurred also below the studied fissure ridges as shown by the lateral rotation (about horizontal axes), in places, of their flanks (De Filippis et al., 2012);
(5) The central feeding conduit of fissure ridges is notable for its strong mineralization both in the Denizli Basin and elsewhere in fissure ridge areas (Bargar, 1978; Selim and Yanik, 2009; Temiz et al., 2009). In contrast, the Tivoli plateau has probably been fed through the active fault zone lying beneath the plateau itself (Facenna et al., 2008; Repoli et al., 2010), but no mineralized conduits (with banded travertine) are so far known across this plateau.
(6) The Denizli and Tivoli travertines have significantly different C- and O-isotope signatures (Fig. 8);
(7) Despite similar annual precipitations, the present water discharge of the Tivoli area is about ten times greater than that of the Denizli Basin.
Fig. 11. Parallel growth models for plateau (left column) vs. fissure ridge (right column) travertines. In both cases, the bedded travertine, which forms the bulk of the deposits, mostly grew when the water table was high and close to the surface (probably warm and/or humid periods). Fluid discharge is identified as the main factor driving the difference between the growth of fissure ridges and plateaus. Such a discharge (i.e., controlling the pore pressure) is probably responsible for the different opening rates of fractures through which the travertines of Tivoli and Denizli were fed (see text). Fluid discharge is most probably modulated by climate oscillations, but the fact that the present discharge of the Tivoli area is about 10 times larger than that of the Denizli area for similar annual precipitations induces us to suppose that, also at the time of travertine deposition, the hydrologic system of Tivoli should have discharged much more fluid than the hydrologic system of Denizli. Moreover, the present fluid discharge of the Tivoli area pertains to a basin significantly smaller than the Denizli Basin (Fig. 2). During cold periods (low-stand of the water table), in contrast, due to a general lowering of the water table, the growth of bedded travertine was substantially inhibited. During these latter times, the Tivoli plateau was affected by partial erosion, whereas the Denizli fissure ridges were injected by veins of banded travertine formed for the periodic upward exsolution of CO₂-rich fluids from a depressed water table.

Fig. 12. The lesser amount of fluid discharge in the Denizli Basin has driven the preferential vertical growth of the travertine deposits (aggradation; Kamara and Akkoy fissure ridges), whereas the larger amount of fluid discharge in the Tivoli area has driven the lateral growth (progradation) of the travertine plateau. Such a large discharge of geothermal fluids, in fact, must have made it possible the precipitation of travertine also far away from the geothermal springs. In the case of the fissure ridges, on the contrary, a large part of travertine must have precipitated right close to the axial springs thus favoring the vertical growth of fissure ridges.
the water discharge at the time of travertine deposition is unknown, the present values of water discharge versus precipitation (i.e., sharply different water discharges for similar precipitations; Table 1) suggest that, also at the time of travertine deposition, the hydrologic system of Tivoli would have discharged much more fluid than the hydrologic system of Denizli. In our model (Figs. 11 and 12), the water discharge is considered the main engine keeping the fractured conduits that fed the travertine deposits open. The higher pressure of water connected with this larger discharge at Tivoli may have activated faults and fractures (i.e., passive tectonics) in the Acque Albune basin, thus promoting a twofold effect: (1) the opening up of pathways for the ascension of large volumes of geothermal fluids and (2) the subsidence of the basin. These processes must have led to a well fed and subsiding geothermal lake where the Tivoli travertine plateau grew for about 100 ka with an associated massive volumetric deposition rate. In particular, the large amount of water discharge caused the precipitation of CaCO3 also far away from the geothermal springs hence driving the lateral progradation of the travertine plateau and ultimately its large volume compared with that of the Denizli fissure ridges.

The Denizli fissure ridges, on the other hand, were fed by reduced volumes of water and, therefore, will have grown under a reduced pressure. Such a reduced pressure led to a slow opening of feeding fissures and the consequent buildup of fissure ridges with alternating precipitation of external bedded travertine and internal bedded travertine during warm and cold periods, respectively (Fig. 11). The slow opening of fissures (Fig. 6) is probably one of the concurrences of fluid discharge driven by the tectonic activity and the fluid discharge, whereas a preferential aggradation occurred at Denizli (Fig. 7a and c). One simple explanation concerning the apparent lack of banded travertine across the Tivoli plateau is the possibility that this type of travertine may indeed be present in the deep and unexposed portion of the travertine plateau. However, in the Denizli Basin, Uysal et al. (2009) interestingly ascribed the growth of banded travertine during cold periods to the CO2 oversaturation of deep reservoirs. Host rock fracturing in response to seismic shaking and fluid overpressure may have resulted in rapid erosion and expansion of the dissolved gas, leading to coseismic hydrothermal outflow (Uysal et al., 2009).

This mechanism is very similar to that proposed by Tuccimei et al. (2006) to explain growth and erosion phases of speleothems from the Colli Albani volcano area (see also Zentmyer et al., 2008; Chiodini et al., 2012; Nishikawa et al., 2012). Also REE (rare Earth elements) data from the banded travertine of Denizli Basin indicate that this travertine formed as a thermogene deposit from rapidly ascending (from deep reservoirs) CO2-rich fluids (Uysal et al., 2009). The apparent absence of banded travertine across the Tivoli plateau may therefore also be explained by the absence of external agents (earthquakes?) able to induce exsolution of CO2 from deep reservoirs during cold stages; but this hypothesis is still highly speculative with the available evidence. On the other hand, the lack of erosional features in the Denizli Basin as marked as those of the Tivoli plateau (Fig. 7a and c) is an interesting conundrum of this study. The inclined attitude of the fissure ridge flanks (Figs. 3 and 4a) may explain why erosion was not accompanied by soil development in the Denizli fissure ridges. This inclined attitude may have promoted a rapid runoff of surface waters, thus inhibiting soil development as well as seepage and karst weathering into the ridges.

In synthesis, this parallel study on thermogene travertines exposed in the Denizli and Tivoli areas has identified three main factors for the origin of the travertines themselves and their differential growth: (1) fluid discharge, (2) water table fluctuations (paleoclimate), and (3) fluid-assisted passive tectonics (in addition to the active one). At the time of travertine formation, interactions and feedbacks among these three factors must have been as follows. The fluid discharge and the water table level below the travertine deposits were conceivably modulated by paleoclimate events. The difference in the rate of fluid discharge, in turn, must have influenced both the opening rates of fractures (i.e., passive tectonics), through which the studied travertines were fed, and the styles of travertine growth. A preferential travertine progradation occurred at Tivoli under an abundant fluid discharge, whereas a preferential aggradation occurred at Denizli under a limited fluid discharge. Abundant fluid discharge must have driven the precipitation of travertine also far away (laterally) from the geothermal springs, whereas a small volume of fluid discharge must have caused the precipitation of travertine close to the springs, thus determining the vertical growth of fissure ridges. Future research should better address the causal relationship between fluid discharge and deformation (i.e., passive opening) across fractures that carry the fluids themselves. Whether this relationship will be ascertained also elsewhere, then, in assessing the long-term
geothermal potential or the liability of geologic sequestration reservoirs, fluid supply and its influencing factors such as climate oscillations (i.e., water table oscillations) should be carefully considered. These factors, in fact, may be the cause of brittle rock deformation and consequent fluid outflow even where active tectonics is absent or weak (e.g., Zoback and Gorelick, 2012).

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Appendix A. Supplementary data

Supplementary data associated with this article can be found in the online version, at http://dx.doi.org/10.1016/j.earsrev.2013.04. 004. These data include Google maps of the most important areas described in this article.

References


