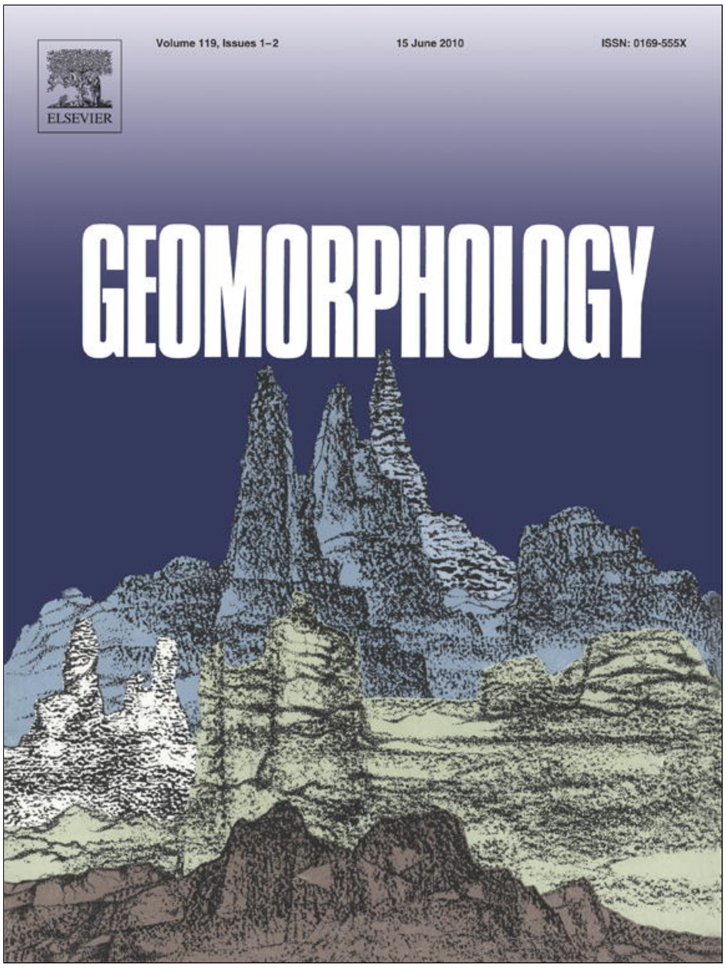
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Volcanogenic origin of cenotes near Mt Gambier, southeastern Australia

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ABSTRACT

The cenotes near Mt Gambier are circular, cliffed, collapse dolines containing water-table lakes up to 125 m deep, floored by large rubble cones. They lie in a flat, coastal plain composed of mid-Tertiary limestone. Most of the deepest cenotes are concentrated in two small areas located along trends sub-parallel to the main joint direction in the limestone. The cenotes do not connect to underwater phreatic passages, and water chemistry data confirm that they are not part of an interconnected karst network. They formed by collapse into large chambers (up to >1 million m^3) that extended 125 m or more below the land surface. Several cenotes have actively growing stromatolites on the sub-vertical walls that started growing at ~8000 years BP.

The caves that collapsed to form the deep Mt Gambier cenotes are much larger than shallow and deep phreatic caves in the area, and do not connect into deep phreatic systems. They were not formed by freshwater/seawater mixing, responsible for many of the well-known Yucatan cenotes, because they are not associated with locations of the mixing zone during previous high sea levels, and are much larger than caves presently forming along the mixing zone near Mt Gambier. Instead dissolution was most likely due to a process whereby acidified groundwater containing large amounts of volcanogenic CO_2 ascended up fractures from the magma chambers that fed the Pleistocene–Holocene volcanic eruptions in the area; deep reservoirs of volcanogenic CO_2 occur nearby.

Cave dissolution could have been due to release of CO_2 during the Mt Gambier eruption ~28,000 years ago, followed by collapse to form cenotes during the low sea levels of the Last Glacial Maximum ~20,000 years ago. The cenotes then flooded ~8000 years ago as sea level rose, and stromatolites began to grow on the walls.

1. Introduction

Cenotes are circular, cliffed, collapse dolines containing water-table lakes. They are spectacular and very popular sites for cave diving, because of their often very clear water and typically large passages (Lewis and Stace, 1980). The best known are those on the Yucatan Peninsula in eastern Mexico (Perry et al., 1995; Smart et al., 2006; Beddows et al., 2007); the term cenote is adapted from the Mayan word 'zenots' for these features (Cardenas, 1996). Cenotes are most common on coastal karst plains with very low topography developed in relatively soft, porous limestone, like the Yucatan Peninsula, Florida Peninsula (Katz et al., 1995), southeastern South Australia (Marker, 1975, 1976; Lewis, 1984; Grimes, 1994), and the Bahamas Banks, where they occur onshore as well as offshore as drowned cenotes, called Blue Holes (Smart et al., 1988a; Smart et al., 1988b; Whitaker et al., 1994). Drowned cenotes have also been recorded on the Great Barrier Reef off the coast of northeastern Australia (Backshall et al., 1979). Cenotes occur in higher elevation limestones in South Africa (Gomes, 1985), northeast Mexico (Gary and Sharp, 2006), and on the Anatolian Plateau in Turkey where they are called “obruk” lakes (Jennings 1985), and they are known from evaporite karst in New Mexico (Land, 2003).

Cenotes, although they are all superficially similar, have a variety of origins, including dissolution by magmatic CO_2 (northeast Mexico; Gary and Sharp, 2006), and collapse into cave systems formed by marine-freshwater mixing solution (e.g. Yucatan; Smart et al., 2006) or by dissolution of evaporite beds (e.g. New Mexico; Land, 2003).

The cenotes in South Australia are well-known as cave diving sites (Lewis and Stace, 1980; Horne, 1993, 1998), but there has been little speculation on how they formed. This paper describes the cenotes and uses their distinctive characteristics to constrain their origin.

2. Geomorphological setting

The South Australian cenotes lie on a broad, low relief, coastal limestone plain around Mt Gambier (Figs. 1 and 2). The plain slopes gently southwards from 30–40 m above sea level (asl) to the sea (Grimes, 1994). To the north is a slightly higher elevation plain (60–

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70 m asl; Fig. 1), also composed of limestone, and partly covered by irregular ill-defined swamps such as Dismal Swamp. These swamps are absent to the south. In the northwest of the area, the inland and coastal limestone plains are separated by the Mount Burr Range, which has volcanic peaks rising to over 200 m asl (Fig. 1). A series of sub-parallel northwest-trending ridges (former coastal sand dunes) cross the limestone plain, rising up to 35 m above its surface.

Apart from the allogenic Glenelg River on the eastern margin, the area has no surface drainage network, due to the flat topography and the porous nature of the limestone (Holmes and Waterhouse, 1983). A small number of short perennial streams issue from coastal springs. Away from the coast, the only surface water within the limestone plain is provided by the cenotes, and some of these were the first points of European settlement in the area.

The limestone plain was originally covered by open eucalypt forest and Casuarina woodland with a few local areas of tussock grassland; this was cleared and converted to pasture following European settlement (Croft et al., 1999).

3. Geological setting

The coastal and inland plains are composed of the latest Eocene to early Middle Miocene Gambier Limestone (Li et al., 2000), which was deposited within the northwest section of the Otway Basin in southern Australia on a cool-water open marine shelf with clear waters (Smith et al., 1995; James et al., 1993). The Gambier Limestone is a bryozoal calcarenite, made up of sand-sized bryozoan fragments and foraminifera cemented by thin rims of calcite around the grains; it has not been deeply buried, and has well-developed inter-granular porosity (James et al., 1993). Abundant bivalves and echinoids occur in some beds; others contain flint nodules, composed of silica derived from sponge spicules. In places the limestone has been dolomitised. The Gambier Limestone rarely contains rare quartz sands, reflecting a restricted terrestrial influence, and towards its base are beds of marly limestone and marl, merging into the underlying Late Eocene Narrawaturk Marl, which is a 15 m-thick glauconitic, fossiliferous, calcareous mudstone (Smith et al., 1995).

The Gambier Limestone is essentially flat lying and thickens to the south from as little as 10 m thick on the Mount Burr Range to 300 m at the coast and 400 m offshore (Fig. 3; Smith et al., 1995). The limestone has been offset along several faults, including the northwest-trending Tartwaup and Nelson Faults (Figs. 1 and 3). Movement on these was probably partly syndepositional; both displace Late Cretaceous and Early Cretaceous strata at depth (SARIG 2009). The faults have also been subsequently reactivated (see below). Nelson Fault is associated with a dolomitised zone (Smith et al., 1995), and has therefore been a conduit for fluid movement. Joints in the limestone trend generally northwest/southeast, sub-parallel to the Tartwaup and Nelson Faults (Marker, 1975; Waterhouse, 1977).

The regression of the sea in the Late Miocene caused some erosion of the exposed Gambier Limestone, so that the age of the limestone on the surface of the limestone plain is generally Early Miocene or Oligocene (Ludbrook, 1971). In the Late Miocene the sea re-advanced and covered much of southeastern Australia (Abele et al., 1988; Holdgate and Gallagher, 2003). As it retreated during the Pliocene and Pleistocene it left behind a series of coastal dunes (former shorelines). In the Mt Gambier area these form northwest-trending ridges of fine- to medium-grained, well-sorted, bioclastic carbonate sand cemented by calcite (Bridgewater Formation; White, 1994). The dunes rise up to 35 m high above the limestone plain and extend sub-parallel to the current coastline for tens to hundreds of kilometers (Fig. 1). The dune ridges are typically several kilometers apart, and each represents a strandline deposited during a specific Pleistocene sea level highstand; between the ridges are thin deposits of shallow marine Bridgewater Formation (Belperio, 1995). The oldest, most northerly Bridgewater Formation dunes (East Naracoorte Range, ~50 km northeast of the Mt Burr Range) are probably about one million years old (Huntley and Prescott, 2001); the youngest, most southerly strandline in the Mt Gambier area (MacDonnell Dune; Fig. 1) is ~125,000 years old, i.e. Last Interglacial (Huntley et al., 1993; Murray-Wallace et al., 1996).

A period of active deformation affected southeastern Australia in the Late Miocene–Early Pliocene (Sandiford et al., 2004; Gardner et al. 2009), due to a change in relative plate motions between the Australian and Pacific plates (Sandiford, 2003). To the east of Mt Gambier, in western Victoria, there were fault displacements of up to 180 m (Paine et al., 2004). It is likely that this tectonism uplifted the limestone plain to the north of the Tartwaup Fault by 30–40 m, formed the uplifted fault block of the Mt Burr Range (Fig. 1), and gently folded the limestone northwest of Mt Schank (Fig. 2).

Pleistocene–Holocene volcanic activity in the Mt Gambier region formed two distinct groups of volcanoes. The older volcanoes lie on the uplifted block of the Mt Burr Range 20–40 km northwest of Mt Gambier (Fig. 1), and comprise 14 eruptive centres associated with three northwest-trending lineaments (Sheard, 1995). The lava flows and ash deposits are overlain by Bridgewater Formation dunes that are lateral equivalents of the Burleigh and Caveton Dunes (Fig. 1), dated as ~240,000 and 320,000 years old, respectively (Huntley et al., 1993; Murray-Wallace et al., 1996). The eruptions therefore occurred prior to 320,000 years ago, and the extensive weathering of the volcanics indicates that they could be older (?mid-Pleistocene; Sheard, 1995).

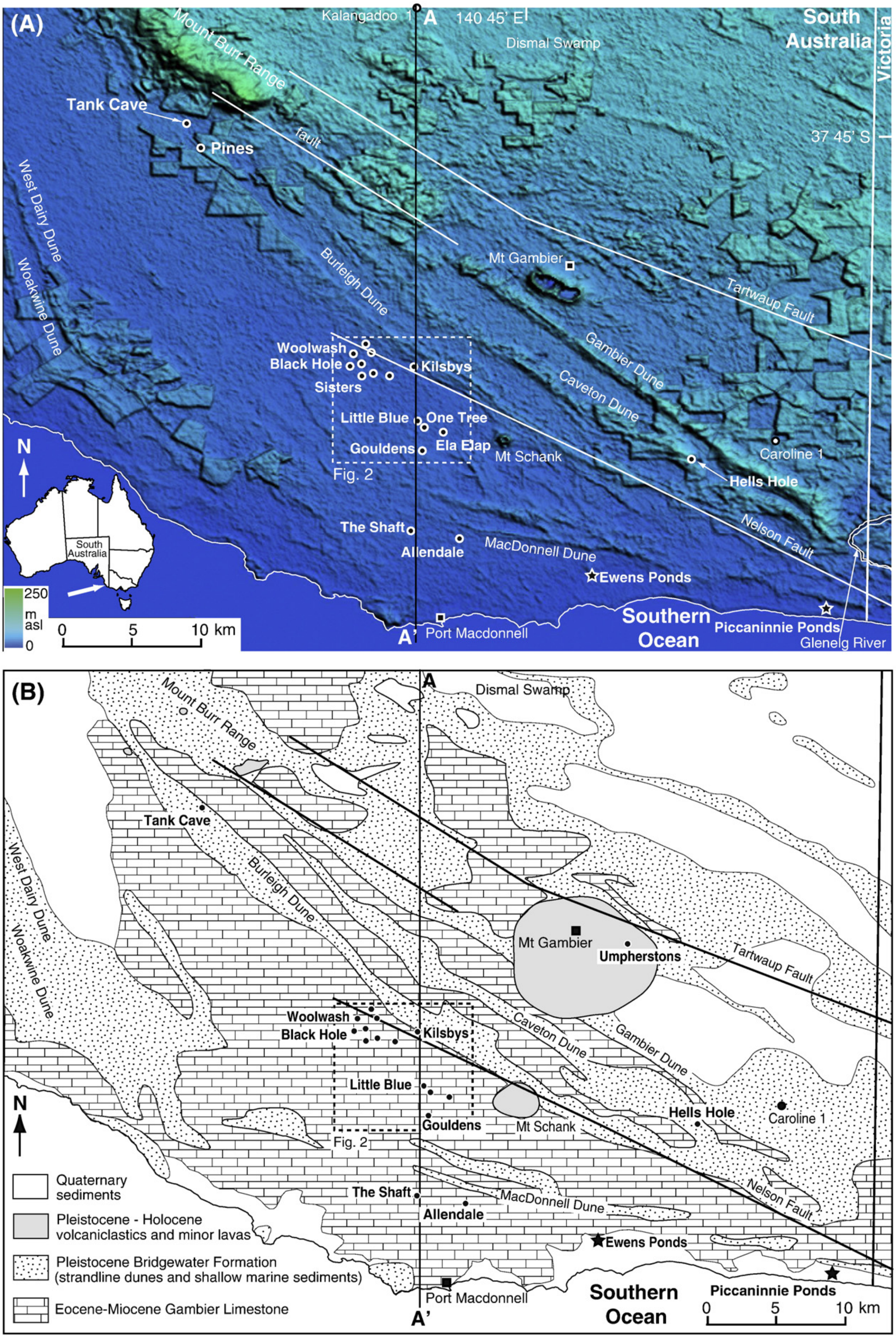
There are also two younger Late Pleistocene–Holocene volcanoes: Mt Gambier itself and Mount Schank (Fig. 1). Mount Schank lies close to Nelson Fault. Both volcanoes are basaltic and are composed largely of scoria (Smith et al., 1995). Four crater lakes are associated with the Mount Gambier complex; the largest, Blue Lake, has a depth of 70 m and represents the local watertable. The crater of Mt Schank is dry. Mt Gambier erupted at ~28,000 years BP (Leaney et al., 1995) and Mt Schank at ~5000 BP (Smith and Prescott, 1987); there may have been a second eruption of Mt Gambier at around the same time (Sherwood et al., 2004).

Where the limestone plain is not covered by Bridgewater Formation dunes or volcanic deposits, it carries thin red-brown (terra rossa) soils except in the swampy areas, where sandy podzols are present (Blackburn, 1983). South of Mt Gambier are extensive areas of limestone pavement where the soils have been stripped. These pavements show rounded exhumed sub-soil karren, with flat, shallow solution pans and small-scale rillenkarren (Grimes, 1994). There is some induration of the weathered limestone surface, often enough to obscure the bedding.

4. Climate

The climate of the Mt Gambier region is classified as Mediterranean or Csb in the modified Köppen system. It is characterised by moist cool winters and hot dry summers. Rainfall in the area is 700–800 mm, and decreases to the north. Mean daily maximum and minimum temperatures range from 22–24 °C and 12–14 °C in January to 13–14 °C and 5–7 °C in July (respectively). Potential evaporation exceeds rainfall from

Fig. 1. (A) Digital elevation model of the Mt Gambier area, showing location of major cenotes and caves (filled circles), springs (stars), main towns (filled squares), boreholes Kalangadoo 1 and Caroline 1 and cross-section line (Fig. 3). Blocky appearance of some areas is due to the presence of tree plantations. DEM downloaded from Shuttle Radar Topographic Mission website. (B) Geological map of the Mt Gambier area, covering the same area as the digital elevation model; modified from Sprigg et al. (1951). The location and extent of Tartwaup Fault are well constrained from stratigraphic bore logs and top of Early Cretaceous Eumeralla Formation and top of Late Cretaceous Sherbrooke Formation seismic horizons, all from SARIG (2009). Nelson Fault is shorter than Tartwaup Fault and less well constrained; there are additional relatively short faults in this area on the seismic maps that could not be confirmed as affecting the Gambier Limestone. The unnamed fault marks the southwestern edge of the uplifted block of the Mt Burr Range.


(A)
Kalangadoo 1
A
140 45' E
South Australia
Victoria
Dismal Swamp
Mount Burr Range
Tank Cave
Pines
fault
37 45' S
West Dairy Dune
Woakwine Dune
Burleigh Dune
Mt Gambier
Tartwaup Fault
Gambier Dune
Caveton Dune
Woolwash
Black Hole
Kilsbys
Sisters
Little Blue
One Tree
Gouldens
Ela Elap
Mt Schank
Fig. 2
Caroline 1
Hells Hole
Nelson Fault
The Shaft
Allendale
MacDonnell Dune
Ewens Ponds
N
South Australia
250
m
asl
0
0
5
10 km
A'
Port Macdonnell
Southern Ocean
Piccaninnie Ponds
Glenelg River
(B)
A
Dismal Swamp
Mount Burr Range
Tank Cave
West Dairy Dune
Woakwine Dune
Burleigh Dune
Mt Gambier
Umpherstons
Tartwaup Fault
Caveton Dune
Gambier Dune
Woolwash
Black Hole
Kilsbys
Little Blue
Gouldens
Mt Schank
Hells Hole
Caroline 1
Fig. 2
N
Nelson Fault
The Shaft
Allendale
MacDonnell Dune
Ewens Ponds
Quaternary sediments
Pleistocene - Holocene volcaniclastics and minor lavas
Pleistocene Bridgewater Formation (strandline dunes and shallow marine sediments)
Eocene-Miocene Gambier Limestone
Port Macdonnell
A'
Southern Ocean
Piccaninnie Ponds
0
5
10 km

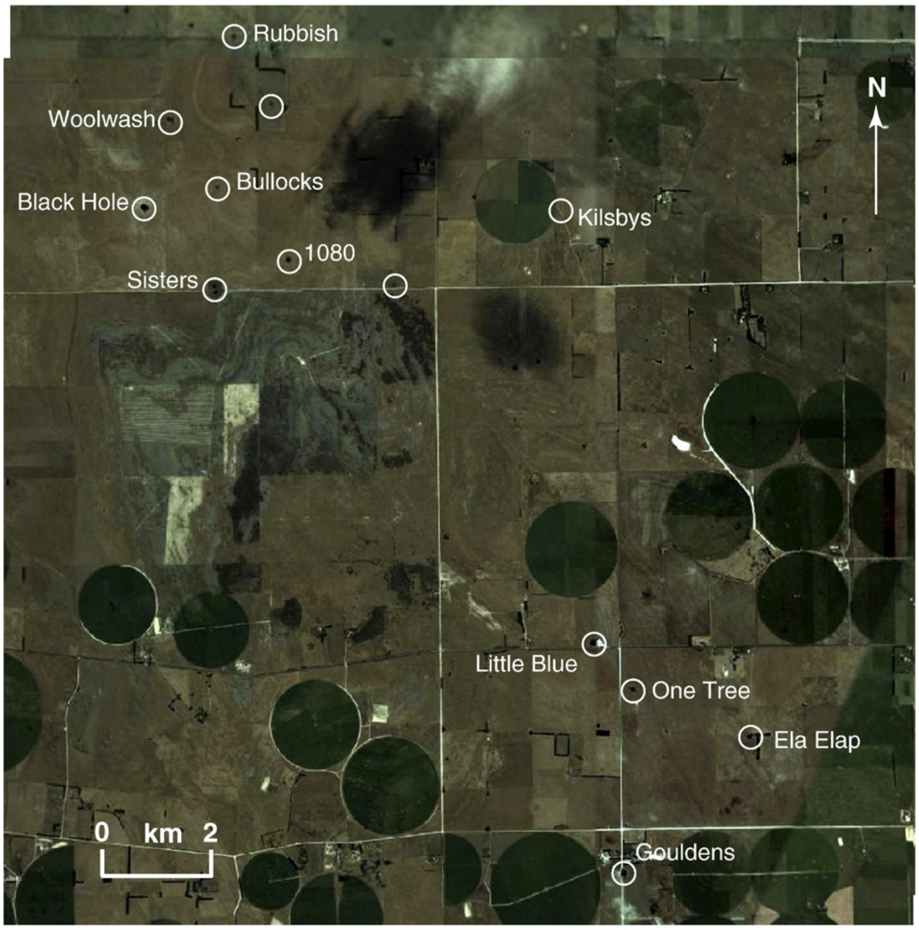


Fig. 2. Main groups of cenotes northwest of Mt Schank; see Fig. 1 for location. Google Earth (DigitalGlobe) satellite image; downloaded 17 April 2009. Note bare limestone pavement south of Sisters and 1080 cenotes, showing broad, low amplitude folding. Large circles are centre pivot irrigation areas.

October to April; the excess precipitation is ~300 mm for the winter months (Waterhouse, 1977; Love et al., 1993).

5. Hydrogeology

The limestone in the Mt Gambier region forms a porous, high-yielding, largely unconfined aquifer that overlies a calcareous mudstone aquitard. The groundwater is extensively used for irrigation and town water supplies because of its high quality (TDS generally 300–600 mg/l) and the lack of surface water resources.

The watertable in the limestone is well-defined (Fig. 3). On the coastal limestone plain it slopes gently towards the coast at a gradient of ~1:1300, so that depth to the water table decreases southwards from 10–30 m near the Tartwaup Fault to intersect the ground surface as springs and swamps along the coast. Along the Tartwaup Fault separating the coastal and inland limestone plains, the hydraulic gradient is substantially steeper (1:40–1:150).

The Gambier Limestone is a dual porosity/permeability aquifer, as it contains both primary intergranular porosity and secondary karstic conduits; the latter are an integral part of groundwater flow throughout the aquifer (Webb and Lithco, 2001). Hydraulic conductivity varies from 2–30 m/day in parts of the aquifer dominated by intergranular groundwater flow to 130–270 m/day in regions with karst fissure development; velocities in the conduits feeding large springs can be of the order of cm/sec (Waterhouse, 1977; Love et al., 1993; Emmett and Telfer, 1994).

A potentiometric low in the water table, the Ewens Ponds–Mount Schank Trough (Waterhouse, 1977), extends across the plain south of Mt Gambier, and includes the highest concentration of karst features (cenotes and springs) in the area, probably due to a greater concentration of conduit flow.

Infiltration of rainfall to the watertable is very high (100–250 mm annually; Waterhouse, 1977; Allison and Hughes, 1978), due to the thin soils and bare rock pavements of porous limestone with open joints. This represents 15–35% of rainfall, compared to an average of <2% across southeastern Australia (Smith, 1998). Most recharge occurs in winter, when rainfall exceeds evaporation, as shown by groundwater stable isotope data (Love et al., 1992; Love et al., 1993). The regional groundwater flow in the limestone aquifer originates on the uplifted plain north of Tartwaup Fault, and has a higher salinity than further south (550–650 mg/l), due to evaporation in the swamps there. As this regional groundwater flows southwards to the coast, it is diluted by low salinity recharge infiltrating rapidly through the limestone, reducing the salinity of the groundwater to 300–550 mg/l in the central part of the plain (Webb and Lithco, 2001).

Along the coastline, a seawater wedge intrudes several kilometers inland (Fig. 3); the gradient and location of the freshwater/seawater interface can be constrained from the gradient of the watertable (according to the Ghyben–Herzberg relationship; Domenico and Schwartz, 1990). Geophysical transects across the area (King and Dodds, 2002) show that the interface is stepped rather having a uniform slope, but the overall gradient is confirmed by the salinity profile from a coastal borehole adjacent to the spring at Piccaninnie Ponds, where the fresh/saline interface is at about 120 m depth. The shaft at Piccaninnie Ponds extends down ~110 m and brings to the surface somewhat saline water (~1300 mg/l) from the underlying seawater wedge (Webb and Lithco, 2001).

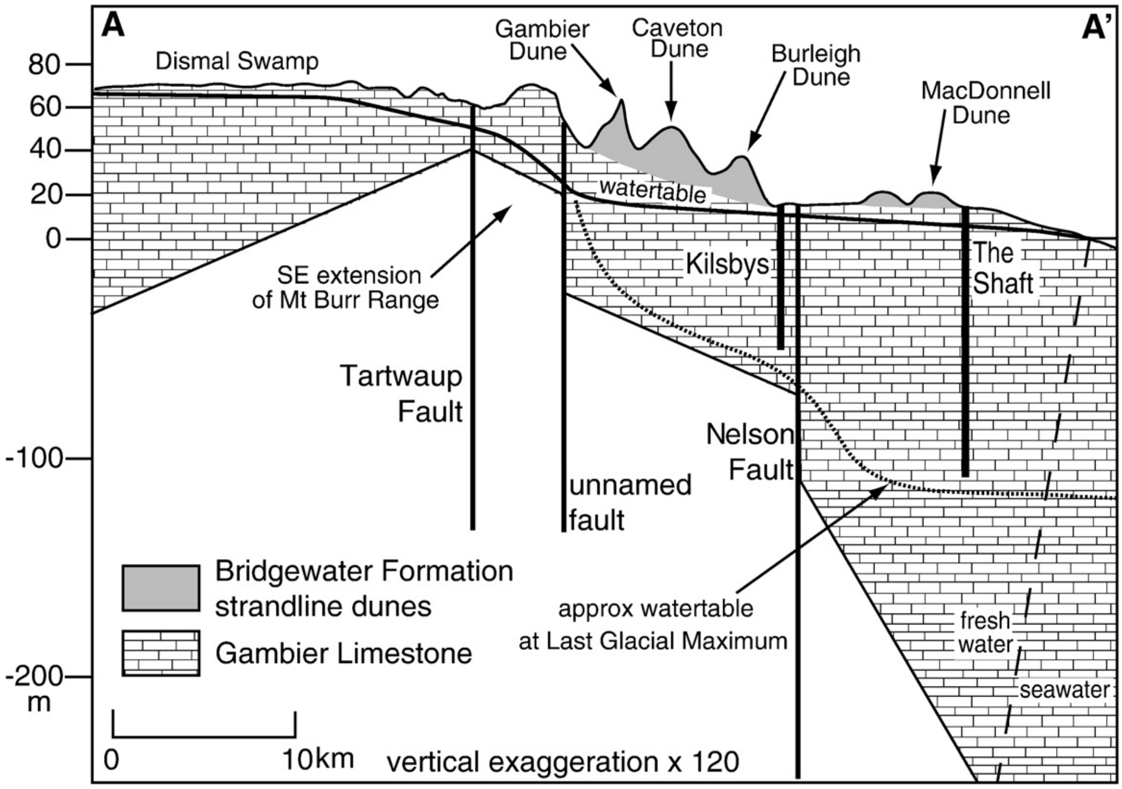


Fig. 3. Geological cross-section; see Fig. 1 for location. Constructed using stratigraphic bore logs and top of Early Cretaceous Eumeralla Formation and top of Late Cretaceous Sherbrooke Formation seismic horizons, all from SARIG (2009). There are additional relatively short faults shown in this area on the seismic maps, but they could not be confirmed as affecting the Gambier Limestone. The likely landward extent of the freshwater/seawater interface at present is shown (from King and Dodds, 2002), as well as the position of the watertable at ~20 ka during the Last Glacial Maximum. Note that thin deposits of shallow marine Bridgewater Formation between the dune ridges, particularly north of Gambier Dune, have been omitted for clarity.

6. Karst features

6.1. Cenotes

The coastal limestone plain is punctured by about 20 cenotes (Figs. 1 and 2), circular cliffed dolines which intersect the water table (Figs. 4A–C and 5B) and contain lakes 20 m to ~125 m deep. The majority of the cenotes, including most of the deepest and volumetrically largest ones, are concentrated in two small areas 12–15 km from the coast (Fig. 1), each ~4 km in diameter, located 5 km west and 10 km northwest of Mt Schank (Fig. 2). Scattered through these main areas of cenotes are several shallow dry collapse dolines; these appear to represent deep cenotes that have been almost completely filled in by collapse, so the floor has risen above the watertable. These dry cenotes have been included in the following discussions of cenote distribution. Collapse dolines are also found in the Mt Gambier area associated with shallow phreatic caves (see below) and some contain shallow lakes, but these are believed to have a separate origin from the deep cenotes, as discussed later. In the following descriptions and discussions, the term 'Mt Gambier cenotes' is used sensu stricto to refer to those containing deep lakes.

The lakes within the cenotes range in size from <100 m^2 and a few meters in diameter (e.g. Bullocks) to ~3000 m^2 and 65 m by 45 m (Black Hole; Figs. 4C and 8), and are surrounded by generally vertical to overhanging walls up to 12 m above the water (Figs. 4A–C and 6), but some cenotes have been degraded by erosion, modifying the sheer vertical walls to gentler slopes (e.g. Woolwash). Several feature 19th century ramps cut through the walls to provide stock access to water. Hells Hole (Fig. 5B) has walls ~30 m high, but only the bottom half of this is Gambier Limestone; the top 10–12 m is cut through the Caveton Dune (Fig. 1), which is 320,000 years old (Murray–Wallace et al., 1996). Below the vertical walls the cenotes widen into flooded collapse domes with central large cone-shaped accumulations of talus covered by silt (Fig. 6). The sediment on the cenote floors contains only small amounts of organic material, because of the sparse vegetation on the surrounding limestone plain (Fig. 4C). The talus cones are up to 60 m high; inclined collapse passages may extend down their sides for up to a hundred meters or more, before pinching out. The extensive collapse passage in Black Hole is aligned northwest–southeast (Fig. 8). The lakes in the cenotes generally submerge the talus cones and are shallowest in the centre (a few meters to 30 m), rapidly deepening down the sides to reach total depths from 20 m (Woolwash) to ~65 m (Black Hole; Fig. 8; Kilsbys cenote, Fig. 10) to ~125 m in The Shaft (Fig. 7). The Shaft is a very large flooded collapse dome without a circular cliffed doline opening to the surface (i.e. a proto-cenote); the present entrance is a small vertical solution tube (Fig. 7). Most of the cenotes appear to be quite stable. However, collapse within them has not completely ceased, as evidenced by the collapse of the Allendale cenote. This opened to the surface in the mid-1800's; repeated attempts to fill it in were finally abandoned in 1971, the road was diverted around it and the hole was fenced off.

The Mt Gambier cenotes are entirely collapse chambers (Fig. 6); from their size it is possible to estimate the volume of the original cave that has collapsed, assuming that the rubble occupies a 75% greater volume than solid limestone (typical bulking factors for the volume increase from solid rock to rubble are 75% for sandstone and limestone; Berkman, 1995). Because post-collapse solution may have removed a proportion of the rubble, any estimate will be a minimum. It is also possible to approximately estimate the maximum depth of the original cave from the calculated original volume, by assuming that the dimensions of the base of the rubble cone do not change with depth. Black Hole contains a rubble pile approximately 60 m high, 70 m wide and 160 m long; the top of the rubble cone lies ~10 m below the surface

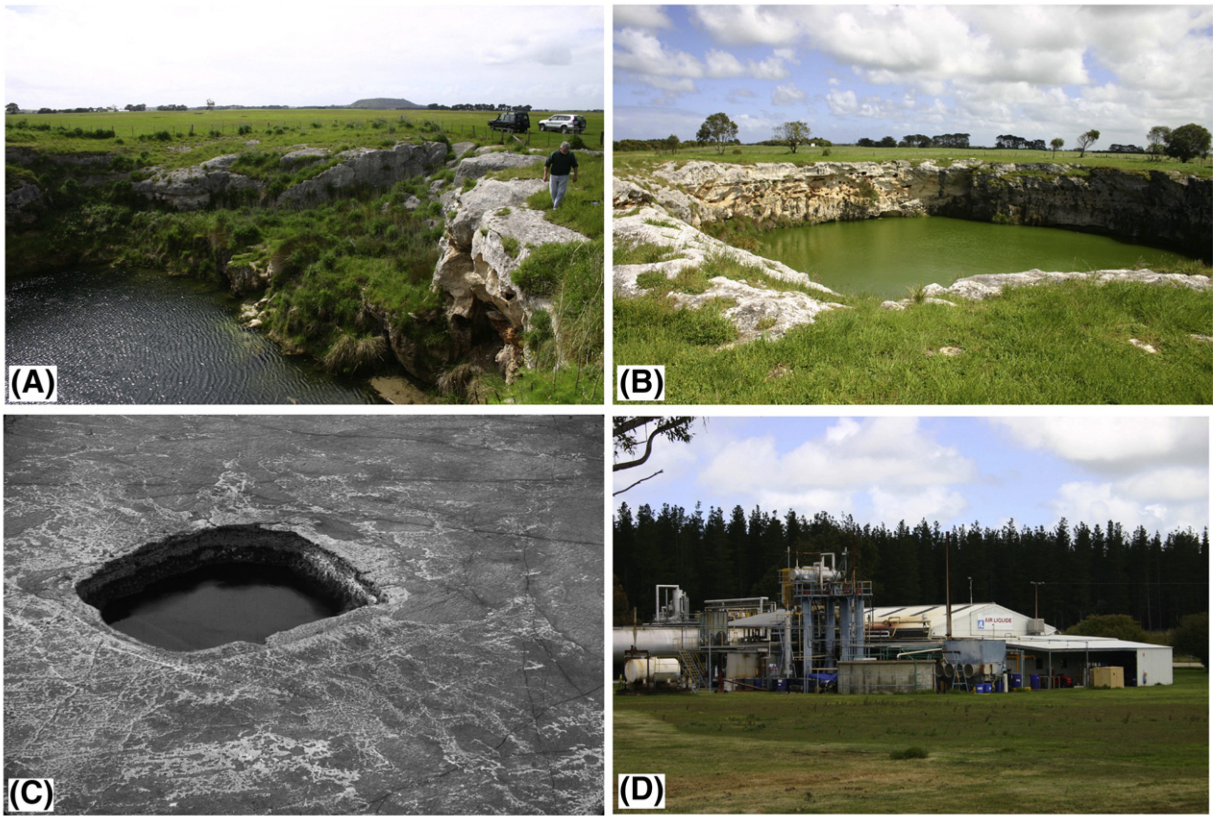


Fig. 4. (A) One Tree cenote; note Mt Schank volcano on the horizon. (B) Little Blue cenote, note small pits and cavities in the exposed limestone wall of the cenote. (C) Black Hole cenote from the air; long diameter of cenote is ~65 m; photo courtesy of Ian Lewis. (D) CO_2 processing plant at Caroline 1 well. See Figs. 1 and 2 for locations. Photos by John Webb unless indicated otherwise.

(Fig. 8). Therefore the cave that collapsed to form this cenote was originally very large (>1 million m^3) and extended up to 110 m below the present land surface (Fig. 3). The Shaft is 125 m deep and the original chamber was also very large (~1.5 million m^3) and could have extended down to ~150 m below the surface. Obviously these estimations of the original size and depth of the caves are only general indications, but the dimensions can nevertheless be used to constrain the origin of the cave (discussed further below).

A spatial analysis of the cenotes (including dry filled cenotes around Mt Schank) shows that they apparently lie along the

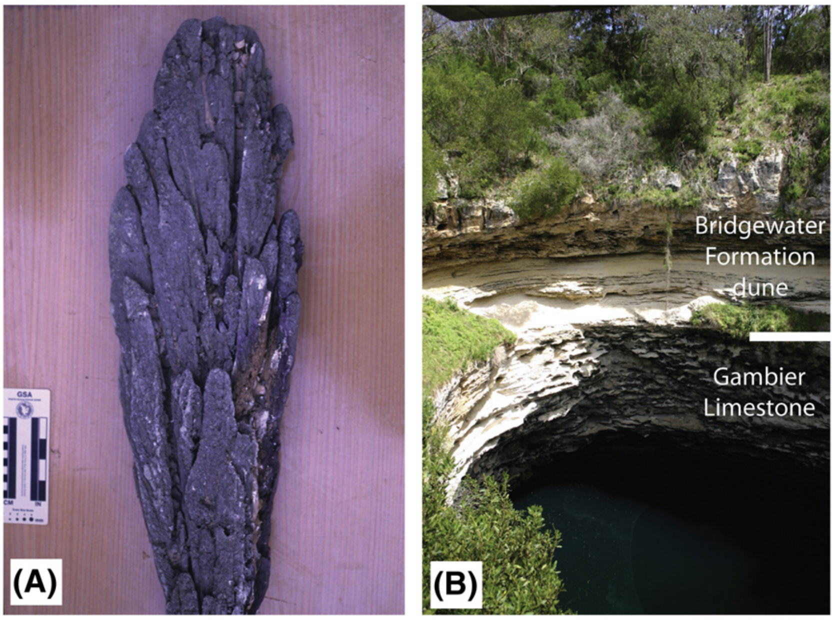


Fig. 5. (A) Columnar stromatolite from Gouldens Hole; scale bar in inches (right) and cm (left). (B) Hells Hole cenote, showing how collapse has penetrated the 320,000 years old Caveton Dune. See Figs. 1 and 2 for locations. Photos by John Webb.

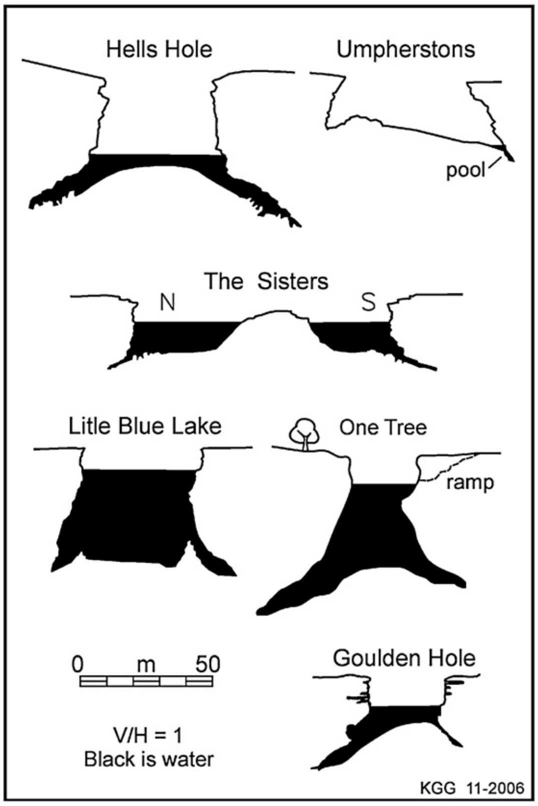


Fig. 6. Cross-sections of representative cenotes. See Figs. 1 and 2 for locations.

intersections between lines trending 050° and 320° (Fig. 9), and in fact the walls of Kilsbys cenote are joint planes with these orientations (Fig. 10). The latter direction parallels the axial traces of low amplitude folds within the Gambier Limestone northwest of Mt Schank (Fig. 2). It is also the dominant joint direction in the Gambier Limestone in this area (Marker, 1975), and is sub-parallel to the Nelson and Tartwaup Faults (Fig. 1) and the regional stress pattern associated with the formation of the Otway Basin. In fact ~25 km northwest of Mt Gambier, a line of elongated dolines follows the surface trace of the Tartwaup Fault for over 2 km (Grimes, 1994).

Diving in the cenote lakes has failed to reveal any phreatic cave passages; the walls of the cenotes beneath the watertable show exclusively collapse features. One cenote (Gouldens Hole) has a small associated phreatic cave above the water table; this is a single, narrow, 0.5 m diameter cylindrical passage extending horizontally for ~40 m from the southeastern side. This is the only known example of a phreatic passage intersected by a cenote.

In the limestone walls above almost all of the cenote lakes there are intensely-perforated zones consisting of numerous small horizontal solution tubes up to 0.5 m in diameter, frequently concentrated in particular beds (Fig. 4B); these also occur for a few meters below water level. These tubes are surface weathering features, as they do not penetrate more than a meter or two into the cenote walls, as shown by exposures cut through the limestone at some of the cenotes.

Water chemistry and stable isotope data from the cenote lakes (Allison, 1974; Webb and Lithco, 2001) show that the groundwater within them is almost always supersaturated with respect to calcite, so the cenotes are not being actively dissolved at the moment. Although many of the cenote lakes become thermally stratified over summer, there is little chemical stratification, and bottom waters have

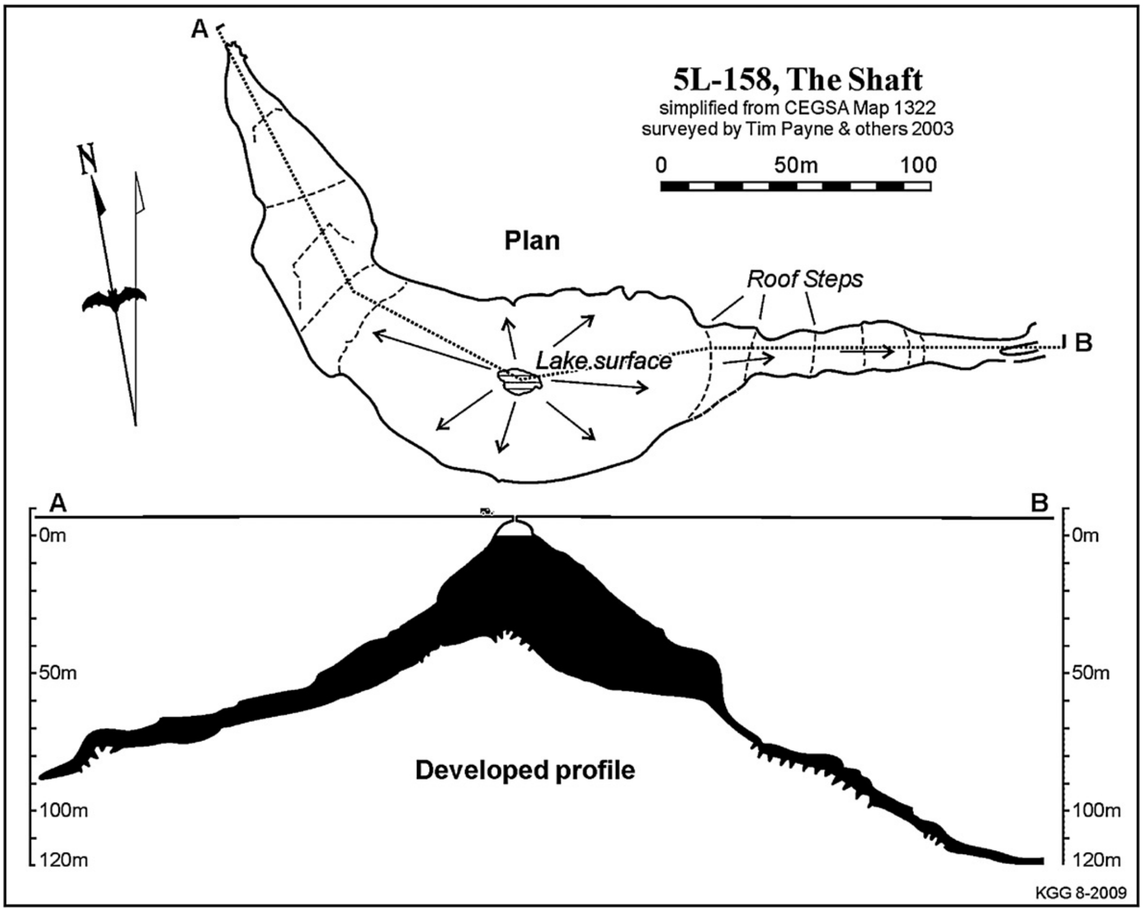


Fig. 7. Map of The Shaft. See Fig. 1 for location. Additional data provided by T. Payne and C. Brown.

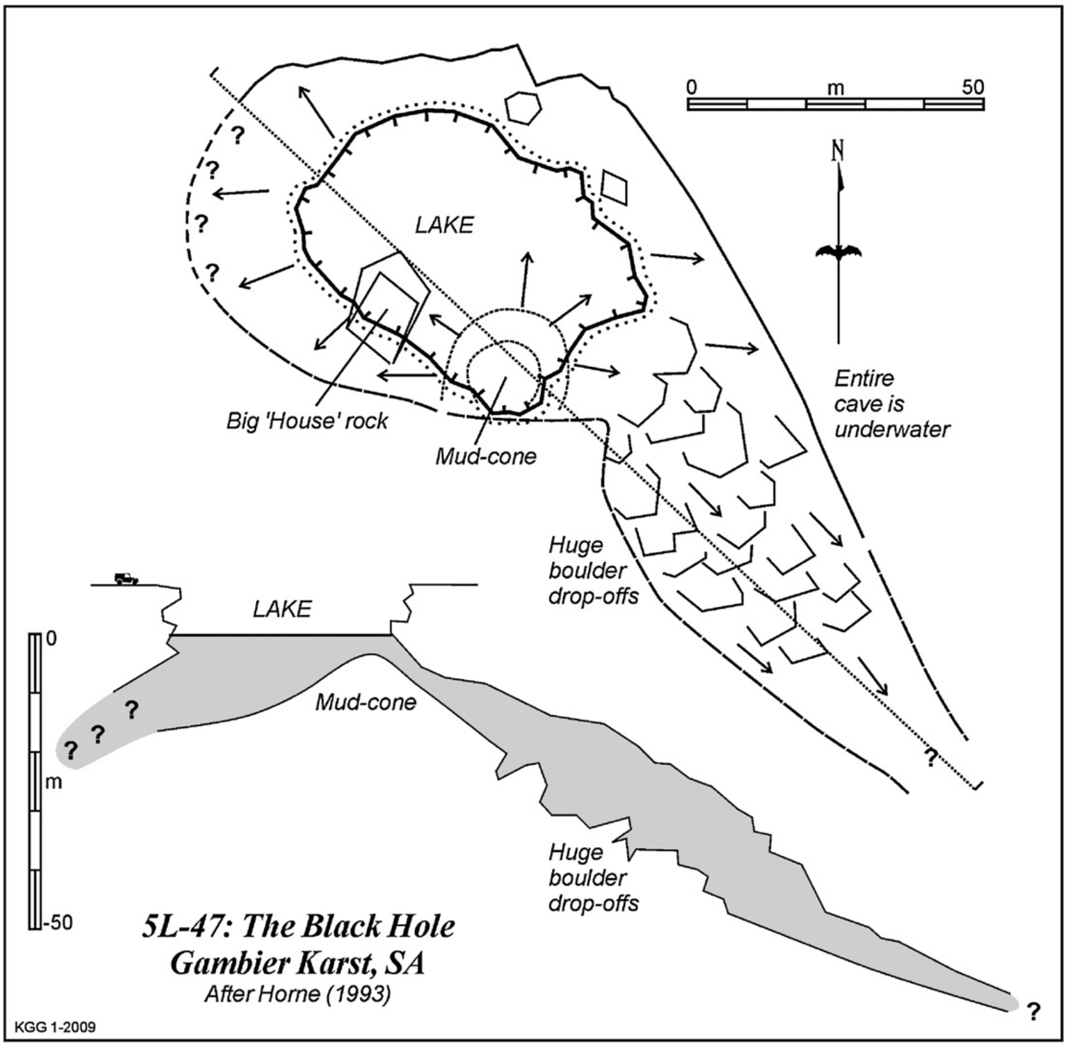


Fig. 8. Map of Black Hole. See Figs. 1 and 2 for location.

virtually the same composition as surface waters. The water in the cenote lakes is a combination of dilute local surface recharge of rainwater through joints and small shallow conduits, along with a variable component of more saline regional groundwater that is supplied by porous inter-granular flow. Differences in the relative contributions of these two components result in substantial variations in cenote water chemistry (Webb and Lithco, 2001), even between nearby lakes such as those in the Sisters cenote, which have different colours and water compositions despite being only 30 m apart. Compared to Little Blue Lake and Gouldens Hole, which are fed largely by relatively saline regional groundwater as porous inter-granular flow, many of the other cenote lakes are less saline, indicating direct local input of dilute rainwater. Overall, the variability indicates that the cenotes are not part of an interconnected karst fissure network, even though many of them lie both along major joints in the limestone and within a potentiometric low in the water table (Ewens Ponds-Mount Schank Trough), which reflects a region of joint-controlled conduit flow.

The only speleothems (stalactites, stalagmites or flowstone) in the deep cenotes are 10 cm stalactites submerged at a depth of 22 m in 1080 cenote (Lewis 1984). However, most' have actively growing stromatolites on the subvertical walls down to a depth of 25 m (Thurgate, 1995, 1996; Horne, 1998). The stromatolites are platy and columnar structures (Fig. 5A); the columns are up to 0.5 m across and several meters long, growing diagonally upwards towards the light on the cenote walls, and are composed of laminated, very fine-grained calcite precipitated by the microbial communities growing on their surfaces. Dating of the base of a 2 m-long columnar stromatolite from a water depth of 8 m in Black Hole shows that it started growing at ~8000 years BP (Table 1; Kelly, 1998), which represents the time when that level of the cenote was flooded as sea level rose after the Last Glacial Maximum.

The $^{87}Sr/^{86}Sr$ ratio of the calcite of the stromatolites averages close to 0.7088 (Table 1), slightly greater than the isotopic signature of the

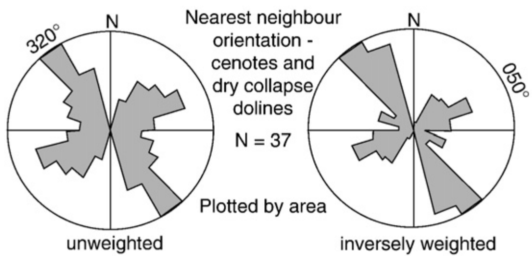


Fig. 9. Rose diagrams of direction to nearest neighbour for cenotes (containing lakes) and collapse dolines (dry) in the Mt Gambier area ($N=37$). Pairs closer together than 150 m are ignored, because they may represent the same collapse chamber at depth. Weighting is by inverse of distance between neighbours; this reduces the influence of isolated dolines and cenotes and emphasises the directions found between closely-spaced features in the main cluster. The diagrams are plotted with the area of each segment, rather than the radius, proportional to the summed value in each direction, as recommended by Cheeney (1983). Cave entrances were excluded from this analysis as many are in Bridgewater Formation dune limestone and are hence unrelated to the caves in the Gambier Limestone.

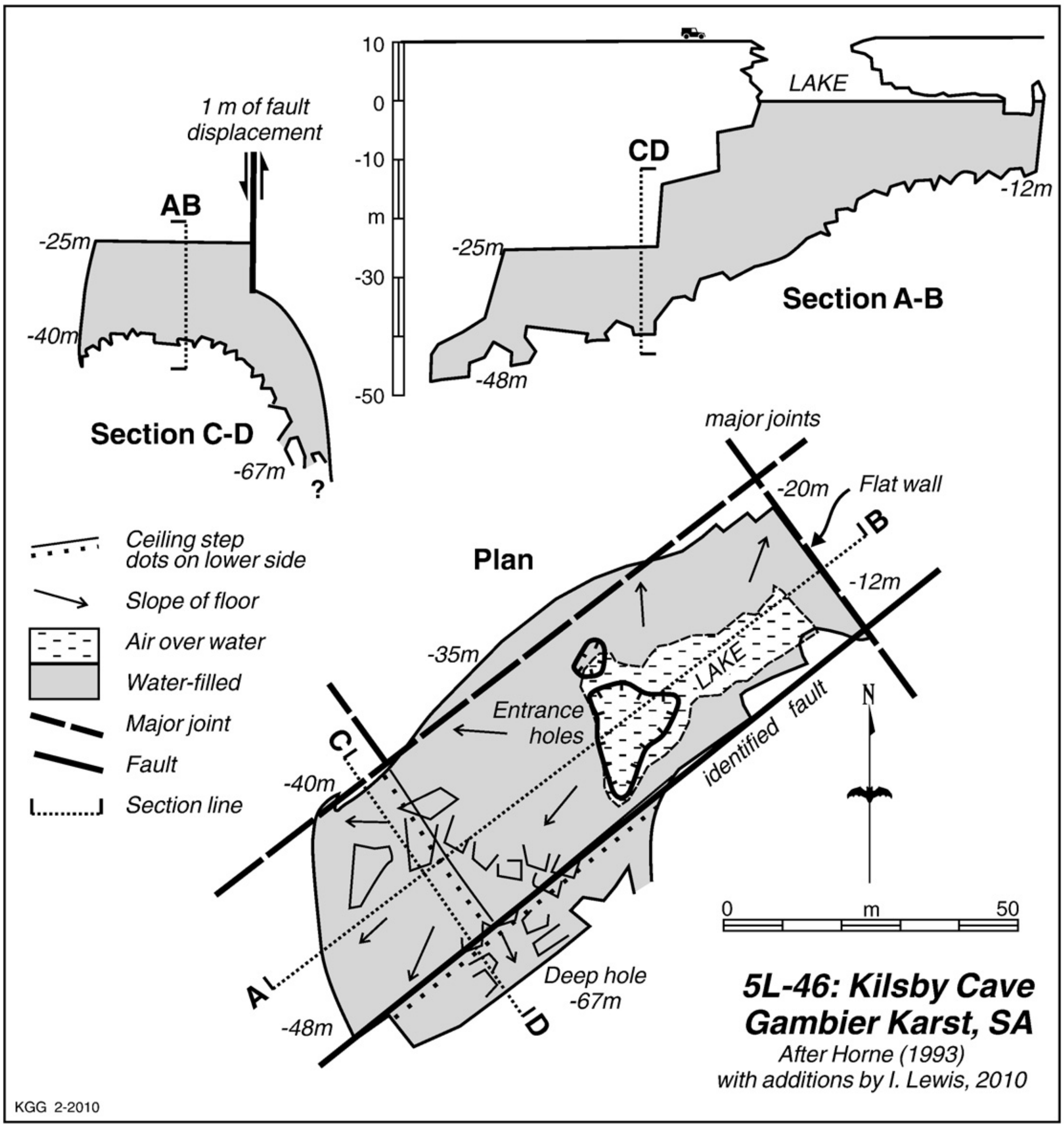


Fig. 10. Map of Kilsbys. See Figs. 1 and 2 for location.

host Gambier Limestone (0.7083, corresponding to the Early Miocene age of the limestone at this location; Howarth and McArthur, 1997). The increase above the limestone value reflects the input of strontium in rainwater (~0.1 ppm), which has a Sr isotope ratio of ~0.7095 near the coast in western Victoria, close to the seawater value of 0.70917 (Raiber et al., 2009). However, one stromatolite sample has a substantially lower ratio (0.7079; Table 1), probably due to a volcanic influence; the Quaternary volcanics of the region have ratios of 0.7037–0.7058; Price et al., 1997). The anomalous sample is bracketed by ^{14}C dates of 4410 and 6930 cal. years BP (Table 1; Kelly 1998), and probably reflects the input of strontium from ash that fell into the lake during the nearby Mt Schank eruption, which occurred at ~5000 years BP (Smith and Prescott, 1987).

6.2. Phreatic caves

Phreatic caves are scattered around Mt Gambier; over 50 are known. They are mostly shallow, horizontal, joint-controlled systems running northwest–southeast (Lewis, 1984; Horne, 1993; Grimes, 1994). Passages are either narrow vertical fissures or oval phreatic tubes with solutionally sculpted walls. The walls in some caves show incuts and undercuts left by old watertables. In contrast to the cenotes, speleothems are often present, although they are generally not abundant. Because of the proximity of the water table to the surface, many caves have water-filled sections. Tank Cave (Fig. 11), the longest, has over 7 km of mostly underwater passages as an extensive maze of phreatic tubes 2–15 m wide and 2–3 m high, with a few larger chambers, and extends to a maximum depth of <30 m.

Many of the primary phreatic passages have been modified by breakdown to form collapse domes and rubble-filled passages. The domes occasionally extend to the surface as dry collapse dolines. These dolines are generally not very deep so they are dry or contain shallow watertable lakes (e.g. Fossil Cave, Fig. 11), and they typically connect to shallow phreatic caves. Pines cenote is ~30 m deep and most likely formed by collapse into a phreatic cave, as it lies near Tank Cave (Fig. 1) and its depth is close to the maximum depth of Tank Cave. This cenote may have collapsed due to a strong local earthquake in 1897; one of the roof blocks in the lake still has a tree stump embedded in it (Lewis, 1984).

Spatial analysis of the longest phreatic cave, Tank Cave, shows that passage orientation is dominantly along a trend aligned at 320° (Fig. 11), reflecting the regional joint structure.

In addition to the shallow caves described above, geophysical logs indicated the presence of a cavity at a depth of 160.4–165.6 m in

Table 1
^{14}C dates (from Kelly 1998) and Sr isotopic analyses of columnar stromatolite from Black Hole; anomalous Sr isotopic analysis indicated in bold.

Distance from base of stromatolite (mm)	$^{87}Sr/^{86}Sr$ ratio[a]	AMS ^{14}C dating data[b]			
		PMC	$\delta^{13}C$ (‰PDB)	Calibrated age (yrs BP)	Uncertainty ($\pm 1\sigma$)
68	0.70880				
59	0.70880				
51	0.70882				
42.5	0.70876				
40	0.70875	42.62	−5.6	4413	−110+25
34	0.70879				
30	0.70879				
28	**0.70789**				
22		32.63	−5.9	6930	−119+81
19	0.70882				
13	0.70886				
7	0.70874				
1	0.70874	26.25	−6.0	8606	−74+100

[a] Analytical procedure: powdered samples mixed with ^{85}Rb–^{84}Sr spike, dissolved and prepared using Sr. Spec columns, loaded on single Ta filaments with phosphoric acid, run in static multicollection mode on Finnigan-MAT 262; mass fractionation corrected by normalizing to $^{86}Sr/^{88}Sr = 0.1194$. Within-run precisions for $^{87}Sr/^{86}Sr$: $\leq\pm 0.00004$ (2SE mean), maximum error for unknowns: ± 0.00007. Blank corrections are insignificant.
[b] Analysed at Antares Mass Spectrometer, Austalian Nuclear Science and Technology Organisation, Sydney; corrected for dead carbon; calibrated using Stuiver et al. (1998).

borehole 7021–3410 (CAR 061) near the coast at Ewens Ponds (K. Mott, pers. comm.).

6.3. Springs

The two major coastal springs are Ewens Ponds and Piccaninnie Ponds (Fig. 1). Ewens Ponds comprises a series of three shallow ponds that feed Eight Mile Creek, with a total discharge of around 2 m^3/s.

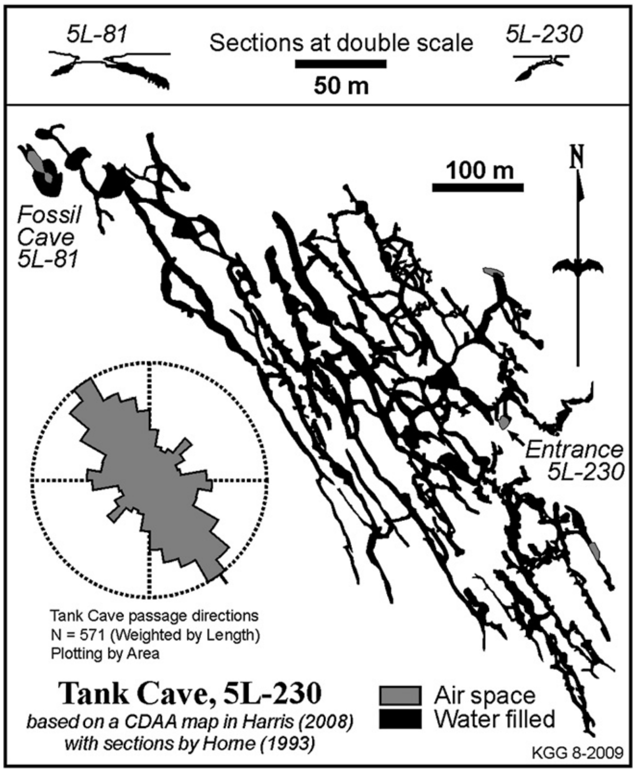


Fig. 11. Map of Tank Cave, the longest phreatic cave in the Mt Gambier area, with spatial analysis of passage directions. See Fig. 1 for location (Harris, 2008).

Piccaninnie Ponds has a discharge of ~1 m^3/s, and contains the spectacular Chasm, a narrow, elongated, funnel-shaped shaft extending to a depth of around 110 m, with well developed phreatic sculpturing on the walls at all depths. Elsewhere springs issue from small dolines, caves or swamps, and some bubble out of the sand on the beach below high tide. Fishermen occasionally report fish schools offshore around what may be undersea freshwater spring outlets.

7. Cenote origin

Because cenotes contain watertable lakes, they are restricted to areas with shallow watertables, mostly very flat coastal limestone plains which rise only tens of meters above sea level. The morphology of most cenotes throughout the world is very similar, i.e. circular cross-section and bell-shaped vertical profile with a central cone of rubble. However, this is just a characteristic of collapse into a large chamber within limestone with low joint density; the formation of the chambers causing the collapse can be due to any of the mechanisms of cave formation. The cenotes on the Yucatan Peninsula are believed to have formed by collapse into cave systems formed by both freshwater/seawater mixing along the coast (Smart et al., 2006) and focused groundwater flow through a high permeability zone developed along ring fractures around the buried Chicxulub impact crater (Perry et al., 1995). The El Zacaton cenote in northeastern Mexico has been dissolved by volcanogenic fluids (Gary and Sharp, 2006), and cenotes in New Mexico have formed by collapse into cavities dissolved in underlying evaporite beds (Land, 2003). All possible modes of origin for the Mt Gambier cenotes will now be discussed, beginning with the least likely.

In this discussion the term 'phreatic' is used in a restricted sense for cave systems dissolved by normal fresh groundwater, and excludes caves dissolved by processes that involve addition of other fluids, e.g. seawater or volcanogenic carbon dioxide.

7.1. Collapse into cavities dissolved in evaporite beds

There are no evaporite units known from the Gambier Limestone or any of the underlying formations (Smith et al., 1995), so formation of the Mt Gambier cenotes is not related to evaporite dissolution.

7.2. Progressive collapse due to dissolution by under-saturated basal waters

In the Bahamas, the basal waters of many of the cenotes are reducing due to the presence of large amounts of organic material (Smart et al., 1988a); the inland cenotes in the Yucatan Peninsula also show this effect (P. Smart, pers. comm.). These waters contain H_2S which oxidises to sulphuric acid, and are often under-saturated with respect to calcite; the adjacent wall rock may be highly corroded, indicating that the cenote is actively enlarging laterally. Thus, these cenotes may be undergoing a process of progressive dissolution, with accompanying collapse as a result of undercutting.

This process is not active in the Mt Gambier cenotes. The basal waters in these are neither reducing nor under-saturated with respect to calcite, and the wall rock at the base of the cenotes shows no evidence of active dissolution.

7.3. Collapse into cavities formed by phreatic processes

Shallow phreatic caves are common in the Mt Gambier area, so potentially the cenotes could have formed by collapse into such systems, as at Fossil Cave (Fig. 11). However, the caves that collapsed to form the cenotes around Mt Schank were much deeper than the shallow systems (at least 125 m below the surface compared to <30 m) and much larger (up to >1 million m^3 compared to only <100,000 m^3 for the most extensive shallow phreatic cave, Tank

Cave). Furthermore, only one cenote (Gouldens) has (accidentally) intersected a shallow phreatic conduit.

It is possible that the cenotes could represent collapses into phreatic systems that formed deep in the Gambier Limestone during times of low sea level. For example, during the Last Glacial Maximum the watertable inland of Kilsbys cenote lay close to the base of the Gambier Limestone (Fig. 3), and phreatic caves could have formed at this level during that period. However, these deep caves would have formed by the same processes as the shallow phreatic caves around Mt Gambier, and would therefore have been similar in size, i.e. too small to form the cenotes. Furthermore, none of the cenotes connect into deeper phreatic systems.

The Yucatan Peninsula cenotes around the Chicxulub impact crater appear to have formed by phreatic dissolution focused along deep fractures associated with the crater, and are clearly aligned along these fractures (Perry et al., 1995). The cenotes around Mt Schank lie within a potentiometric low in the watertable (Waterhouse, 1977), but they do not represent a large-scale trend indicative of deep-seated, focused flow along this trough, as they are concentrated in two relatively small, more or less circular areas (Figs. 1 and 2).

Thus, the Mt Gambier cenotes do not appear to have formed by collapse into shallow or deep phreatic systems.

7.4. Collapse into cavities formed by freshwater/seawater mixing

The cenotes along the east coast of the Yucatan Peninsula are mostly collapse dolines that connect to extensive, flooded networks of passages, developed in a zone 8–15 km inland of the coast (Smart et al., 2006). The sub-parallel anastomosing systems are relatively shallow, deepening from less than 10 m near the coast to ~30 m deep 8 km inland; they run roughly perpendicular to the coast and feed to point discharges. The caves are composed of horizontal elliptical tubes and narrow vertical passages, with generally rounded dissolution morphologies; walls are often pocketed or fretted. The major process of cave development is dissolution along the mixing zone between fresh and saline groundwater, and the modern mixing zone is often coincident with the widest part of the cave passages. Because of the flat topography (inland elevations are only 20 m asl) and extensive karstic permeability, the water table gradient is very low (~1:17,000). This results in the very gradually increasing depth of the freshwater/saline interface into the interior (~30 m deep at 12 km inland according to the Ghyben–Herzberg relationship; Domenico and Schwartz, 1990), and represents the primary control on the depth distribution of the caves. The small hydraulic gradient results in a very slow groundwater flow towards the sea despite the good drainage provided by the large cave systems.

In the Bahama Banks there are a large number of 'blue holes', the entrances to flooded cave systems; the best-studied are on Andros Island (Smart et al., 1988a, 1988b; Whitaker et al. 1994). The coastal blue holes have developed along extensional fractures parallel to the bank margin, and often contain extensive horizontal passages that have dissolved along the mixing zone at the base of the freshwater lens. Some caves are deeper, and formed at a previous lens base below present sea level, associated with the much more extensive freshwater lens that existed during glacio-eustatic emergence of the Bahama Bank (Smart et al., 1988b).

Inland on both the Yucatan Peninsula and Andros Island are scattered cenotes, flooded vertical shafts up to 110 m deep and 50–150 m in diameter, that bell out at depth (Whitaker et al., 1994; Beddows et al., 2007). Some of the cenotes have short horizontal passages leading off, although these are usually inaccessible due to breakdown (as at Mt Gambier). These inland cenotes most likely formed by collapse into deeper cave systems that formed along the mixing zone during times of lower sea level. They may be actively enlarging as a result of basal dissolution and undercutting (described above).

In contrast to the Yucatan and Andros examples described above, the Mt Gambier cenotes are not connected to phreatic cave systems. The freshwater/saltwater interface, which is responsible for the formation of the Yucatan and Andros caves and cenotes, does not extend inland to the Mt Gambier cenotes at present (Fig. 3), because of the configuration of the limestone aquifer, together with the relatively steep hydraulic gradient (~1:1,300). During previous high sea levels, the freshwater/saltwater mixing zone at Mt Gambier would have been located inland, beneath each of the strandline dunes as they were deposited. Therefore, if the cenotes in the main area near Mt Schank were formed by this process, they should lie along trends parallel to and beneath or close to a dune ridge, but there is no association (Figs. 1 and 3).

However, there is no doubt that enhanced dissolution along the freshwater/saltwater interface has helped to form the cave systems that feed the coastal springs south of Mt Gambier. The salinity profile from a coastal borehole adjacent to the spring at Piccaninnie Ponds shows the fresh/saline interface is at about 120 m depth, and this is verified by geophysical data for the area (King and Dodds, 2002). The shaft at Piccaninnie Ponds extends down ~110 m and brings somewhat saline water (~1300 mg/l) to the surface, representing a mixture of freshwater within the aquifer and seawater from the underlying wedge that intrudes inland (Webb and Lithco, 2001). Mixing between fresh and saline water results in under-saturation and carbonate dissolution (e.g. Smart et al., 1988a), and this has probably been responsible for dissolving the deep shaft at Piccaninnie Ponds (which has well-developed solutional fretting on its walls), assisted by the substantial flow rate in the spring (~1 m^3/s). In contrast, none of the main cenotes shows solutional sculpture and they are much larger than the Piccaninnie Ponds shaft, which has a volume of <50,000 m^3.

7.5. Collapse into cavities dissolved by volcanogenic CO_2

Lewis (1984) suggested that volcanogenic CO_2 introduced from depth might be a factor in formation of the Mt Gambier cenotes. Two deep wells drilled 30 km north and 17 km southeast of Mt Gambier (Kalangadoo 1 and Caroline 1 respectively; Fig. 1) intersected liquid CO_2 at depths of 2.1–2.5 km; Caroline 1 is a commercial producer of liquid CO_2 (Fig. 4D; Sheard, 1995). Analyses of $\delta^{13}C$ and isotopes of He, Ne and Xe collectively indicate a primitive mantle source for the CO_2 (Chivas et al., 1987, Caffee et al., 1988, Giggenbach et al., 1991), i.e. it is volcanogenic. Therefore dissolution of the large deep chambers that formed the cenotes could have occurred through acidified groundwater containing large amounts of volcanogenic CO_2 ascending up fractures from deep-seated reservoirs, which were related to the magma chambers that fed the Quaternary volcanoes in the area. The ascending waters probably travelled up fractures in the Gambier Limestone; the cenotes appear to lie along the intersections of major joints (Fig. 9). Deep groundwaters can contain substantial concentrations of dissolved CO_2 because of the high pressures and can therefore dissolve large amounts of limestone (Palmer, 1991). As these groundwaters rise they cool, and this may be enough to maintain their aggressiveness even as the decreasing pressure causes CO_2 to exsolve (Dublyansky, 2000). Furthermore, as the ascending high CO_2 waters mixed with lower CO_2 groundwater in the Gambier Limestone, the mixing effect would have enhanced the solutional aggressiveness and caused even more limestone dissolution.

Caves with a hypogene (deep-seated) origin are known from around the world and can be dissolved by waters acidified by CO_2 and/or oxidation of H_2S; if the waters are hydrothermal in origin, they are also characterised by relatively high temperatures (Dublyansky, 2000). Groundwater in the Mt Gambier area at present does not have elevated temperatures and lacks H_2S or abnormally high sulphate levels (derived from oxidation of H_2S), suggesting that the dominant hypogene process was dissolution by volcanogenic CO_2.

Elsewhere in southeastern Australia there are cold, effervescent springs containing volcanogenic CO_2 but lacking H_2S (Cartwright et al., 2002).

Other cenotes have also been ascribed to volcanogenic karstification. Sistema Zacaton in northeastern Mexico contains a ~300 m deep cenote (El Zacaton); the system lies in a plain close to a volcanic centre and deep groundwaters have elevated levels of H_2S and relatively high temperatures, indicating a direct volcanic influence (Gary and Sharp, 2006). The El Zacaton cenote is larger than the Mt Gambier examples, probably because it experienced a much greater, longer term supply of volcanogenic fluids. There is only a small rubble cone, probably because of removal by ongoing dissolution. In addition, the cenote lies sufficiently far inland that it did not drain during the fall in sea level at the Last Glacial Maximum, which may have caused the collapse in the Mt Gambier cenotes (see below).

8. Timing of deep, volcanogenic cenote formation

The dissolution of the deep caves at Mt Gambier could have been due to a single event when a large amount of volcanogenic CO_2 was released during the mid-Pleistocene(?) Mt Burr eruptions and/or the late Pleistocene Mt Gambier eruption. The strontium isotope composition of the stromatolites, which began growing ~8000 years ago (Table 1), shows no volcanic influence apart from a single sample that probably records ash erupted from Mt Schank. Thus there has been no input of volcanogenic CO_2 for at least the last 8000 years, so dissolution preceded the Holocene Mt Schank eruption.

The collapse of the cenotes may have occurred as a single, relatively recent event; the majority of the cenotes are now quite stable and there is little ongoing breakdown. Collapse at Hells Hole postdates 320,000 years ago, the age of the dune penetrated by the breakdown, and the cenotes were clearly present prior to 8000 years ago when the stromatolites began growing. Much of the collapse in limestone caves generally occurs when the caves are drained and the buoyant support of the roof by water is removed. Thus cenote collapse around Mt Gambier would most likely have occurred during the Last Glacial Maximum (20 ka), when sea level was ~120 m below present sea level (Peltier and Fairbanks, 2006) and the deeper volcanogenic caves would probably have been at least partially drained (Fig. 3). In this case the caves would have been air-filled in their present configuration for only a short period of time, before they were flooded by the rapid post-glacial sea level rise, explaining the almost complete lack of speleothems on the cenote walls (shallow phreatic caves in the area often contain speleothems). If the caves formed during the mid-Pleistocene(?) Mt Burr eruption rather than the late Pleistocene Mt Gambier eruption, then they would have been subjected to several periods of draining and refilling as sea level fluctuated, but there is no evidence of this in terms of notching or other lateral solutional activity on the cenote walls.

Therefore deep cave formation may have occurred due to release of volcanogenic CO_2 during the late Pleistocene Mt Gambier eruption ~28,000 years ago, followed by collapse of these caves to form cenotes during the Last Glacial Maximum at ~20,000 years ago. The cenotes then flooded as sea level rose during the post-glacial transgression, and stromatolites began to grow on the cenote walls at ~8000 years ago.

9. Conclusions

The cenotes near Mt Gambier contain lakes up to 125 m deep floored by large rubble cones, so they formed by collapse into large, deep chambers.

A few deep phreatic caves are known in the Mt Gambier area, but they are too small to represent the caves that collapsed to form the cenotes. Furthermore, the cenotes do not connect into any deep phreatic systems.

Freshwater/seawater mixing, which formed the Bahama Banks and eastern Yucatan cenotes, was not responsible for the Mt Gambier examples, because the latter are not associated with locations of the mixing zone during previous high sea levels. Furthermore, the Mt Gambier cenotes are much larger than caves forming along the mixing zone at present in the Mt Gambier area.

The deep caves that originally lay beneath the Mt Gambier cenotes probably have a volcanogenic origin, similar to the Sistema Zacaton in northeastern Mexico. Dissolution was most likely due to acidified groundwater containing large amounts of volcanogenic CO_2 ascending up fractures from reservoirs which were related to the magma chambers that fed the Pleistocene–Holocene volcanic eruptions in the area. Deep reservoirs of volcanogenic CO_2 are known in the vicinity. The dissolution of the caves could have been due to a single event when a large amount of volcanogenic CO_2 was released, most likely during the late Pleistocene Mt Gambier eruption ~28,000 years ago.

Collapse of the deep caves to form the cenotes probably took place when the caves at least partially drained during the very low sea levels of the Last Glacial Maximum. The almost complete absence of speleothems on the cenote walls and the lack of evidence that the caves drained and refilled as sea level fluctuated, suggests that they do not date back to earlier sea-level lowstands.

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