Setting the stage for the genesis of the giant Bendigo ore system

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Abstract: The Ordovician sediments that host the giant Bendigo saddle reef gold deposits consist of a 3 km-thick sequence of turbiditic sandstones and interbedded siltstones and mudstones. Sedimentological studies suggest the succession formed within a major deep marine channel–levee complex similar to those described from contemporary continental margin to basin plain settings outboard of major river systems (e.g. the Amazon, Mississippi and Congo). Black shales, which are commonly the immediate host rocks to the epigenetic gold reefs, are interpreted to be over-bank deposits or abandoned channel fills, developed adjacent to active channels which were sandstone-dominated and had an incised axial thalweg marked by the coarsest-grained sediments present.

Organic carbon content of the black shales at Bendigo varies from 0.2 to 2 wt%, compared with the grey shales, siltstones and sandstones, which vary from 0.05 to 0.2 wt%. Trace elements fall into two main groups: (a) elements that have a linear relationship with aluminium, and are controlled by the detrital clay content (Sn, Ba, Rb, Li, Cs, Mn, Cr and Ti); and (b) elements that show relationships with both aluminium and organic carbon (V, U, Ni, Zn, Cu, Bi, Pb, Se, Ag and Au) and are controlled by both the clay and organic matter content in the carbonaceous shales. The elements in the second group are enriched in the black shale facies.

The background gold content of the black shales in the drill holes distal from mineralization averages 8.9 ppb, compared with the sandstones with 1.5 ppb. Most of the gold in the shales is present in diagenetic pyrite and marcasite, which laser ablation inductively coupled mass spectrometer (LA-ICPMS) analyses indicate varies from 5 to 3850 ppb and averages 370 ppb Au. The geochemical data suggest that this syngenetic gold was most likely sourced by erosion of the hinterland, and transported attached to detrital clay particles or as colloidal gold, by a high-volume feeder river system. High Rb/K ratios in the shales support a highly weathered source typical of a giant river system. By analogy with modern systems, following transport into deep marine channel–levee complexes via continental margin canyons, gold and other redox sensitive trace elements were ultimately trapped by reduction, adsorption and complexation with organic matter in the sub-oxic to anoxic over-bank deposits. Oxidation of much of the organic matter during diagenesis released the gold and certain trace elements (Ni, Co, Se, Ag, Cu, Bi, Pb), which became incorporated into diagenetic pyrite. Enrichment of gold in diagenetic pyrite of the black shale facies of the Ordovician turbidites at Bendigo was the first stage in a two-stage process that produced the world-class quartz–gold saddle reef deposits.

Supplementary material: Whole rock analyses for sedimentary rocks in drill holes NBD005 and NBD186, Kangaroo Flat Mine, Bendigo, are available at http://www.geolsoc.org.uk/SUP18732

The goldfields of central Victoria are the world’s premier example of turbidite hosted gold-bearing quartz vein systems and Bendigo is the biggest field, having produced 529 tonnes of vein-hosted gold (Willman & Wilkinson 1992). It occurs as native gold within classical quartz saddle reefs and associated structures developed during the formation of anticlinal fold hinges within the Ordovician turbidite succession of the Bendigo Zone in western Victoria (Fig. 1). The source of the gold in the Bendigo field and other Victorian goldfields has been proposed to be the Cambrian mafic volcanic sequences that occur below the Ordovician succession (e.g. Crawford & Keays 1987; Keays 1987; Bierlein et al. 1998). However, there is a clear association of gold reefs with black shales containing gold–arsenic-bearing diagenetic pyrite, and an alternative view is that the black shales in the host turbidite succession were the source of gold (Wood & Large 2007; Thomas et al. 2011).

The sedimentary setting of the Bendigo Zone turbidites has generally been interpreted in terms of deep marine submarine fan models. However, advances in deep water imaging and sampling over the last 15 years have facilitated the recognition of deep water canyon–sinuous channel–levee systems, which has led to a paradigm shift in our understanding of the nature of sediment transport and deposition in deep marine settings. The relevance of these systems to the Bendigo host rocks with regard to correlating stratigraphic elements at the mine scale has been emphasized by Boucher et al.
In this study, we apply the developing deep marine facies models to the sedimentology and geochemistry of the Ordovician sequence at Bendigo, and investigate their significance for the genesis of the mineralization.

Although there have been numerous studies of the lithogeochemistry of hydrothermal alteration related to the Victorian gold deposits (e.g. Gao & Kwak 1997; Bierlein et al. 1998; Li et al. 1998; Dugdale et al. 2006; Arne et al. 2008), little has been written about variations in the background geochemistry and trace element composition of the host sedimentary rocks. This is probably because the gold ores are epigenetic in nature, and the chemistry of the sedimentary rocks has not been considered relevant to the genesis of the ores and/or the discovery of new deposits.

In this contribution, we present new data on the sedimentology and lithogeochemistry of the host sedimentary rocks, principally black shales, but also grey shales, siltstones and sandstones. The principal aims of this study are to characterize the sedimentary environment and trace element geochemistry of the shales, to discuss the redox environment of shale deposition, and to assess whether the sedimentology and geochemistry of the host rocks has played a role in the genesis of the gold deposits.

**Geological setting**

The host rocks to the Bendigo Goldfields in the Bendigo Zone in western Victoria (Fig. 1) are a c. 3 km thick Lower and Middle Ordovician succession of mass flow deposited sandstone and mudstone, with lesser suspension deposited mudstone and shale termed the Castlemaine Group (Fergusson & Vandenberg 2003). Interpretation of deep seismic data indicates that the Ordovician succession is floored by Cambrian ocean crust (Cayley et al. 2011). Together with the contemporaneous Adamina Group in the Taberabbera and Omeo zones in eastern Victoria (Fig. 1), these form the ‘Ordovician mud pile’ (Vandenberg et al. 2000), which records Late Cambrian to Late Ordovician basin plain sedimentation outboard of the Gondwanan continental landmass (e.g. Cas 1983). The current...
across-strike extent of this deep water succession is >500 km (Fig. 1), and the pervasive development of chevron folds during the Late Ordovician to Early Silurian east–west contractual deformation of the Benambran Orogeny has resulted in shortening of 50–60% (Fergusson & Vandenberg 2003).

In western Victoria, the turbidite pile is exposed in two adjacent structural elements separated by the north–south-trending Avoca Fault (Fig. 1). The western Stawell Zone comprises a Cambro-Ordovician succession termed the St Arnaud Group, and the eastern Bendigo Zone comprises a Lower and Middle Ordovician succession termed the Castlemaine Group. Provenance data indicate that the sediment in both successions was derived from convergent tectonic activity that affected the adjacent Gondwanan continental margin: the Delamarian (Fig. 1) and Ross Orogenies in South Australia and Antarctica, respectively (Gray & Webb 1995; Turner et al. 1996; Fergusson & Tye 1999). Specifically, isotopic and geochronological data indicate that the erosional products of initial Late Cambrian Delamarian folding and broad uplift were deposited as the St Arnaud Group; succeeding rapid exhumation and cooling in the Early Ordovician (Turner et al. 1996) sourced the basal sandstone-rich sediments of the Castlemaine Group termed the Romsey Sub-Group (Vandenberg et al. 2000). Tectonic activity on the Gondwanan margin then appears to have waned, with the succeeding Early and Middle Ordovician Castlemaine Group sediments comprising a broadly upward fining succession.

The most comprehensive study of the sedimentology of the Castlemaine Group is that of Cas et al. (1988). The sandstone beds that characterize the Orodovician succession have been widely interpreted as turbidites owing to their broadly lensoidal geometry, and the range and arrangement of primary sedimentary structures that are consistent with the Bouma sequence (Cas et al. 1988). The lack of subsequent reworking and preservation of intervening fine-grained suspension deposits indicates that they record an entirely deep water environment on the basin plain outboard of the continental rise and slope. Cas et al. (1988) defined four sedimentary facies: amalgamated sandstone, thick-bedded sandstone, mixed sandstone–mudstone and mudstone. Three additional facies have been added in the most recent consideration of the sedimentology of the Ordovician of Victoria (Fergusson & Vandenberg 2003): massive sandstone–granulestone, black shale and chert.

The base of the Castlemaine Group succession, dominated by the thick-bedded sandstone and mixed sandstone–mudstone facies, and intervals of the massive sandstone–granulestone facies are locally present. Beds have sheet-like geometries consistent with a flood of clastic detritus from the most rapid phase of exhumation during the Delamarian Orogeny (Turner et al. 1996). In the remainder of the Castlemaine Group, the sandstone facies are arranged into lensoidal bodies up to 70 m thick that may show upward thinning, upward thickening or symmetrical bed thickness trends (Cas et al. 1988). They have generally been interpreted in terms of base of slope submarine fan models (Mutti & Ricci-Lucchi 1978; Normark 1978), with many considered to represent broad channel fills on the basis of features such as scoured sharp bases and mudstone intraclasts. Others, presumably those with upward-thickening bedding trends, have been interpreted as constructional lobes.

Cas et al. (1988) noted that, while such sand-filled channels are consistent with mid fan regions, no inner fan facies are preserved. In addition there is no overall upward-coarsening trend indicative of fan progradation, rather the reverse is true, and the overall Castlemaine Group succession is gradually upward fining. This led to the conclusion either that the system represented a very large submarine fan, the extent of which was much greater than that of the Bendigo Zone (c. 20 000 km² without taking into account the effects of deformation), or that there was no fan present, and the upper part of the Castlemaine Group succession represented channelized sedimentation on the basin plain environment.

In a regional sense, one aspect of the sedimentology of the Castlemaine Group that stands out is the thickness and abundance of graptolitic black shales. In the turbidite sequences to the west and east such black shales are relatively rare and fine-grained sediments are more commonly cherts that have faunal assemblages consistent with a truly pelagic origin (Cas et al. 1988). It has been suggested that the Castlemaine Group was deposited in an area of deposition that was the site of oceanic upwelling that favoured organic productivity, while anoxic bottom conditions favoured organic preservation (Fergusson & Vandenberg 2003).

**Deformation and mineralization in the Bendigo Zone**

The Bendigo Zone is characterized by classical upright open to tight chevron folds with wavelengths of 150–500 m indicating relatively simple east–west compression during the Benambran Orogeny (Gray 1997; Fergusson & Vandenberg 2003). Narrow zones of stronger deformation and cleavage development occur in the hanging walls of the four major north–south-trending high-angle reverse faults that subdivide the zone, and are interpreted as a linked system of imbricates that splay off
a mid crustal detachment zone (Cox et al. 1991a, b; Gray & Willman 1991; Fig. 2). Overall, the Castlemaine Group and underlying Cambrian oceanic crust have been shortened by up to 70% above regional detachment that occurs at depths of 15–17 km (Gray 1997), and as a result the succession has been thickened from c. 6 to c. 20 km. The Mt William Fault at the eastern edge of the Bendigo Zone represents the exhumation of the detachment zone (Fig. 1).

The Bendigo Goldfield is hosted by symmetric and persistent north–south-trending chevron folds that have controlled the geometry of the gold reefs (e.g. Vandenberg et al. 2000; Figs 2 & 3). The eastern edge of the field is sharply bounded by the more deformed succession in the hanging wall of the west-dipping Whitelaw Fault, but there is a gradational decline in grade and abundance of mineralized structures to the west (Willman & Wilkinson 1992; Fig. 2). Gold reefs are localized on west- and east-dipping reverse faults that truncate fold hinges, and saddle reefs controlled by thick competent sandstone beds. These range from simple folded, bedding concordant extensional veins to complex fault/vein structures (Willman & Wilkinson 1992). Historically, most of the gold came from the hinge zones and eastern limbs of three anticlines, and best mineralized areas correspond to domal culminations in the fold axes; however, west-dipping fault and associated extensional veins also host significant gold. Alteration haloes around the veins and reefs comprise disseminated pyrite and arsenopyrite up to several tens of metres away from lodes (Thomas et al. 2011) and more widespread carbonate poikiloblasts (Cox et al. 1991b), both of which have been used as vectors to ore. The most common fluid inclusions in the mineralized veins are low-salinity, two-phase, water-rich and may contain CO₂ and/or CH₄ (Cox et al. 1991b), and galena-sphalerite thermometry indicates fluid temperatures around 300 °C. Ar/Ar dating of sericite from an extensional vein gave ages of 439 ± 2 and c. 420 Ma (Foster et al. 1998).

![Fig. 2. Schematic cross section through the central part of the Bendigo Zone in the region of the Castlemaine Goldfield, 40 km south of Bendigo (modified after Willman 1995).](image)

![Fig. 3. Mine scale cross section of the Garden Gully Anticline at Bendigo Mine (section ND1).](image)
In general, the gold mineralization is thought to have occurred during crustal heating associated with the Benambran Orogeny, with the source of the gold being the Cambrian mafic volcanics in the area of the regional detachment (e.g. Keays 1987; Bierlein et al. 1998). The regional location of the Bendigo field in the hanging wall of Whitelaw Fault has been widely used to infer that this was the plumbing system linking the deep Cambrian source in the region of the detachment with the shallower Ordovician succession that hosts the mineralized reefs (e.g. Willman 2007; Fig. 2). In this model, in the area of the Goldfields, ascending fluids were focused up plunge into anticlinal fold hinges in fold culmination zones (Fig. 3), where low-permeability black shales allowed the development of high fluid pressures and therefore hydraulic fracturing and vein formation (Cox et al. 1991b). Black shales are also important on an outcrop scale, as reefs and veins are consistently highest in grade where they crosscut these units. Cox et al. (1991b) argue that this is because they were the source of CH4-rich fluids that migrated into adjacent fault zones when pressure was decreased during failure events, where they mixed with more oxidized auriferous fluids, thereby promoting Au deposition.

**Sedimentology**

When Cas et al. (1988) hypothesized that the Castle-maine Group may represent channelized sedimentation in the basin and plain environment, little was known about the sedimentary processes that operated in these abyssal depths. However, subsequent advances in detailed bathymetric imaging and the results of deep sea drilling programmes have resulted in the recognition that large-scale sinuous channel systems are, in fact, the major conduits for the transfer of sediment from the continental shelf across the slope and rise to the deep ocean. The current state of knowledge of these systems has been recently reviewed by Wynn et al. (2007). Most documented examples are from large passive margin fans supplied by major rivers (e.g. the Amazon, Mississippi and Congo) that supply huge volumes of mainly fine-grained sediment to the deep ocean via shelf margin canyons; however examples have also been recognized from tectonically active margins. In detail, bathymetric imaging indicates that these deep water channels have a wide range of geometries and internal architectures. Owing to the paucity of systematically spaced deep drill cores, investigation of their internal facies patterns has been restricted to seismic interpretations and surface sediment sampling, and no coherent sedimentary facies model has been proposed. Some generalization can however be made, and Figure 4 is a schematic idealized cross section based on the references cited herein.

The channel and associated levee systems commonly have significant positive relief (tens to 100 m) and may extend down-slope from the continental rise for hundreds of kilometres. Although a network of channel–levee systems commonly exist, the channels are subject to frequent abandonment and typically only one channel is active at a time. Abandoned channels are filled with fine-grained deposits that have overspilled from the adjacent active channel.

The main channel can be several kilometres wide; however, it often contains a relatively narrow incised channel (thalweg), which in the case of the Rhone channel is 200–800 m wide and 40–50 m deep (Bellaiche et al. 1981; O’Connell et al. 1991, 1995). Flows within the thalweg are confined (e.g. Zaire channel; Babonneau et al., 2004) and therefore likely to be relatively dense sediment–water suspensions (sandy debris flows). Such dense flows are typically overridden by less dense, turbulent sediment–water suspensions (e.g. Talling et al. 2004) likely to spill over the thalweg margin and deposit sandy turbidites on the main channel floor. Marginal levees are deposited, in turn, by flows over-spilling the main channel and are generally composed of fine-grained turbidites (e.g. Zaire channel, Droz et al. 2003; Indus fan Wynn et al. 2007).

A sedimentary facies log of c. 200 m of stratigraphy that hosts the Bendigo gold reefs is shown in Figure 5. Four of the sedimentary facies previously

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**Fig. 4.** Schematic facies diagram of a deep water channel–levee complex.
Fig. 5. Composite sedimentary facies log of the Ordovician succession at Bendigo Mine.
recognized in the Ordovician of the Bendigo Zone and discussed above are present: massive sandstone–granulestone, amalgamated sandstone, thick-bedded sandstone and mixed sandstone–mudstone. These are described in detail in Cas et al. (1988) but brief summaries of the facies as they occur in the section logged from Bendigo mine follow.

The mixed sandstone–mudstone facies represents the ambient background within which all of the more sandstone-dominated facies were emplaced (Fig. 5). Intervals of the facies are characteristically dominated by massive to laminated, organic-rich and locally pyritic siltstone and mudstone, with varying proportions of intercalated fine-grained sandstone beds. The latter are generally thin (<10 cm) and normally graded, and some preserve ripple marks (Bouma C divisions), all of which are consistent with them representing classical turbidites (e.g. Shanmugam 1997).

Individual beds within the thick-bedded sandstone facies (Fig. 5) are generally less than 1 m thick and comprise massive or normally graded fine- to medium-grained sandstone. Bed bases are usually sharp and planar on the scale of the width of the drill core, but in some cases have scours, lodes and flame structures and some beds have laminated tops (Bouma B divisions), all of which are consistent with this facies also recording classical turbidites.

The amalgamated sandstone facies (Fig. 5) comprises intervals of massive to diffusely laminated fine- to medium-grained sandstone. Contacts between beds are sometimes indicated by thin finer-grained sandstone to siltstone intervals but are of difficult to pick; however, bed thickness seems to range between 1 and 3 m. Mudstone intraclasts are sometimes present that occur within beds, not at bed bases, indicating that they were deposited from sandy debris flows (e.g. Shanmugam 1997; Talling et al. 2004) or sustained hyperconcentrated flows (Kneller & Branney 1995) rather than classical turbidity currents. The massive sandstone–granulestone facies is similar in character but coarser-grained. Mudstone intraclasts up to 1 cm in size are particularly prevalent in the massive granulestones, indicating that these also represent sandy debris flows or sustained hyperconcentrated flows.

Overall, the three thicker bedded facies defined define a series of 13 upward-thinning and fining bedding cycles of between 10 and 90 m in thickness that occur within the ambient background conditions recorded by the mixed sandstone to mudstone facies. The thicker cycles have significant basal intervals comprising the amalgamated sandstone (cycles 1, 4, 5, 10, 11 and 12), or in one case (cycle 6) the massive sandstone–granulestone facies, overlain by intervals of the thick-bedded sandstone facies. The thinner examples (cycles 2, 3, 7, 8 and 9) generally have bases comprising the thick-bedded sandstone facies and upper parts comprising a gradation back to the mixed sandstone–mudstone facies.

In the light of the discussion above on the architecture of contemporary deep water sinuous channel–levee systems, it is our contention that the facies association present in the Bendigo Goldfield (Fig. 5) is consistent with deposition in a series of overlapping deep water channels of the type depicted schematically in Figure 4. We interpret the massive sandstone–granulestone and amalgamated sandstone (Fig. 5) to represent sandy debris flow or sustained hyperconcentrated sediment water suspensions resulting from constrained flow within a deeply entrenched axial thalweg. The thick-bedded sandstone facies comprises classical turbidites (Shanmugam 2002) deposited on the broader channel floor outboard of the axial thalweg, from turbulent flows sourced from elutriation of material from the confined flows. The mixed sandstone–mudstone facies comprises thinly interbedded turbidic sandstones and hemi-pelagic grey to black mudstones deposited as Bouma E division or during subsequent periods of quiescence. Where this facies occurs interbedded with the thick-bedded sandstone facies that represent the channel floor (e.g. the interval between the Big Blue and Rowes Shales; Fig. 5), it is interpreted to represent channel margin levee systems (Fig. 4). However relatively thick and fine-grained (i.e. mudstone dominated) intervals of this facies (e.g. the Owen, Ethan and Rowes Shales; Fig. 5) are interpreted to represent the final phase of infilling once a channel has been abandoned. The overall facies architecture of the Bendigo host succession indicates that it is centred on an area dominated by the channel facies throughout the deposition of a series of sequentially active channel–levee systems.

Analytical methods

Sedimentary rock samples consisted of about 10 cm of half drill core, split and analysed for all major elements by XRF at University of Tasmania, and 40 trace elements by 4-acid digest ICP-MS (method ME-MS61) at the ALS-Chemex laboratory in Brisbane. In addition, low-level gold was analysed by first roasting the sample and using an aqua regia digest with an ICP-MS finish. Organic carbon was determined by subtraction, after measuring total carbon by Leco and inorganic carbon by strong acid dissolution at the ALS-Chemex Laboratories. Results are presented in the Supplementary Material.

The LA-ICPMS analyses employed a New Wave 213 nm solid-state laser microprobe coupled to an Agilent 4500 quadrupole ICPMS, housed at the
CODES LA-ICPMS facility at the University of Tasmania. The methods used are described in detail in Large et al. (2009), Thomas et al. (2011) and Danyushhevsky et al. (2011) and summarized briefly below.

The laser microprobe is equipped with a small volume ablation cell (c. 2.5 cm$^2$) characterized by $<1$ s response time and $<2$ s washout time. Ablation was performed in pure He (0.71 m$^{-1}$). The aerosol carried by the He gas was immediately mixed with Ar (1.23 l m$^{-2}$) after leaving the cell, passed through a pulse-homogenizing device then directly introduced to the torch. Pyrite analyses were performed by laser-ablating spots 40 μm in diameter, the repetition rate was 5 Hz and laser beam energy was maintained between 4 and 5 J cm$^{-2}$. Analysis time was restricted to 90 s, comprising 30 s background (laser shutter closed) and 60 s analysis (laser shutter open). Acquisition times are as follows: major silicate elements (i.e. Mg, Al, Si, Ca) and S = 0.005 s; trace silicate elements (i.e. Zr, Ba, La, Th, U) and Cd = 0.01 s; Se, Mo and Ag = 0.1 s; Te = 0.2 s; Au = 0.4 s; other elements = 0.02 s. The total sweep time was approximately 1.5 s.

Data reduction took place according to standard methods (Longerich et al. 1996) with Fe the internal standard. The primary calibration standard consisted of a Li borate fused glass of pyrite and/or sphalerite mixture developed in-house (Danyushhevsky et al. 2011) and was analysed twice every 90 min with a 110 μm beam size at 10 Hz to correct for instrument drift.

Imaging of sulphides was performed by ablating a set of parallel lines arranged in a grid over the sample so that the space between the lines was the same as the size of the beam. Beam size was chosen as 15, 25 or 30 μm depending on sample size. The beam was rastered over the lines at the same speed as the beam size (i.e. spot = 25 μm, speed = 25 μm s$^{-1}$), at 10 Hz repetition rate. Therefore, every position in the sample was ablated 10 times, contributing to five consecutive pixels in the image. This results in an unprocessed effective resolution which matches the beam size. A set of 20 elements was chosen for analysis, with acquisition time for the majority of elements set to 0.002 s; exceptions were Se (0.004 s), and Ag, Te and Au (0.04 s). Total sweep time was c. 0.2 s. A 13 s delay was left after each line to allow for cell washout. Background levels and drift were measured on the primary standard before and after every image (Danyushhevsky et al. 2011). Maps were generally generated over a period of 1–2 h where drift in sensitivity is minimal. Occasionally, larger maps were generated and a drift correction was undertaken, with instrument drift considered to be linear between the standards.

**Chemical composition and chemostratigraphy of Bendigo turbidites**

Eighty-nine samples of sandstone, siltstone and shale facies were collected from two drill holes NBD 005 and NBD 186 (Fig. 6). The holes together cut the complete named stratigraphic section at Bendigo from the Jude Shale to the Herchells Unit (Fig. 5), and were deliberately selected to be at the greatest distance possible from known mineralization.

The turbidites have a simple mineralogy of quartz, phengitic illite, albite, minor chlorite and trace pyrite and graphite. Quantitative scanning electron microscopy and X-ray diffraction analysis shows that the illite content ranges from around 5 to 30 wt% in the sandstones, and from 30 to 80 wt% in the siltstones and shales. Albite is concentrated in the sandstones, where it varies from 5 to 20 wt%, whereas the siltstones and shales contain only 1–5 wt%. Chlorite and pyrite content is generally $<5$ wt% and carbonate and graphite content $<2$ wt%.

Although the samples were carefully selected to avoid quartz veining and alteration, a further screening process was applied to ensure that all hydrothermally altered samples related to the epigenetic gold mineralization were eliminated from the dataset. Previous research (e.g. Bierlein et al. 1998; Li et al. 1998; Arne et al. 2008) had demonstrated subtle alteration extending from 10 to 200 m from the gold reefs. The inner zones of sericite alteration carrying minor disseminated hydrothermal pyrite and arsenopyrite are surrounded by ferroan carbonate spotting of the sedimentary rocks. The hydrothermal pyrite and arsenopyrite halo is discriminated in whole-rock analyses by elevated As and As/Al ratios (As/Al $>1.2 \times 10^{-3}$), and the outer carbonate spotting halo is discriminated by elevated Ca, Mn and Sr (Ca $>2$ wt% and/or Mn $>750$ ppm and/or Sr $>150$ ppm; Arne et al., 2008). Samples with any of these chemical characteristics were eliminated from the dataset.

For the purposes of this study, the Bendigo turbidites are classified using an Al$_2$O$_3$ v. organic carbon plot (Fig. 7a). The sandstone and shales exhibit a linear Al$_2$O$_3$–C$_{org}$ trend up to 28 wt% Al$_2$O$_3$ and 0.2 wt% organic carbon. Above 0.2 wt% organic carbon, this positive linear trend is reversed, as the black carbonaceous shales increase in organic carbon, but decrease in Al$_2$O$_3$, accompanied by a rise in SiO$_2$ (Fig. 7b). The quartz-rich sandstones contain 5–15 wt% Al$_2$O$_3$, and less than 0.1 wt% organic carbon, whereas the grey siltstones and shales contain 15–28 wt% Al$_2$O$_3$, and 0.04–0.2 wt% organic carbon. The black carbonaceous shales contain from 0.2 to 2.0 wt% organic carbon and from 12 to 26 wt% Al$_2$O$_3$ (Fig. 7a).
The average $K_2O/Al_2O_3$ ratio for the Bendigo turbidites is 0.25 (Fig. 8a), compared with 0.24 for North American Shale Composite (NASC; Gromet et al. 1984), 0.22 for Wedepohl’s Average Shale (WAS; Wedepohl 1995), 0.19 for Post Archean Average Shale (PASS; Taylor & McLennan 1985), and 0.17 for Global Subducting Sediment (GLOSS; Plank & Langmuir 1998). The $K_2O–Al_2O_3$ trend (Fig. 8a) is curvilinear upwards, showing an increase in $K_2O/Al_2O_3$ ratio from a value similar to that of GLOSS for the sandstones, to higher values similar to NASC and WAS for the shales. This trend probably relates to the proportions of illite, smectite and kaolinite clays in the original sediments, with the data suggesting an increase in the proportion of illite (the dominant clay) with increasing Al$_2$O$_3$ content in the shales.

The Rb/K ratio in the turbidites is fairly constant (Fig. 8b), except for a small number of samples with $K_2O > 5$ wt%. Compared with NASC, WAS, PASS and GLOSS, the Bendigo shales have a high Rb/K ratio, which McLennan et al. (1990) considers to be characteristic of ancient and highly weathered sources.

The negative trend for Na–K (Fig. 8c) relates to the inverse relationship between albite (concentrated in the sandstones) and illite (concentrated in the shales). The low Ca content and lack of a Ca–Al trend in the data are related to the low carbonate content of the turbidites and the resultant lack of any significant Ca–Al-bearing phases.

**Comparison with the ‘average shale’ composition of Wedepohl**

Mean whole rock major and trace element values for the three main turbidite facies, based on the dataset in the Supplementary Material, are shown in Table 1, where they can be compared with the average shale of Wedepohl (1995). This comparison demonstrates that, in terms of major element composition, the Bendigo grey and black shales are very similar to the average shale. The only minor differences shown by the Bendigo shales are lower values of CaO and Na$_2$O, complimented by higher values of K$_2$O. In terms of trace elements, the enrichment factors (Tribovillard et al. 2006) listed in Table 1 highlight that, for the Bendigo black shales, Ag, Zn, As and Au are enriched 3–5 times over average shale, whereas Se, Sb and Bi are enriched 5–8 times. Other trace elements generally show enrichments of 1–3 times. The only significantly depleted trace elements in the Bendigo black shale are Mn and Sr, most likely owing to their low carbonate content and reducing conditions during sedimentation.
In an attempt to understand the controls on particular trace elements in the turbidites at Bendigo, they have been plotted against Al, a measure of the detrital clay content of the original sediments (Figs 9–11). Four distinct patterns are present in the plots that allow the trace elements to be subdivided into the following groups:

**Group 1** – Sn, Ba, Rb, Li, Cs, Mn, Cr and Tl (Fig. 9); these trace elements show good linear correlations with Al across all sedimentary facies, with no additional significant enrichment in the carbonaceous facies. These linear relationships are common in shales (e.g. Tribovillard *et al.* 2006), and indicate the trace elements were introduced with detrital aluminosilicate components.
Group 2 – V, U, Ni, Zn, Cu, Bi, Pb, S, Se, Ag and Au (Fig. 10); trace elements in sandstones and grey siltstones–shales exhibit a positive linear correlation with Al, as in group 1. However trace elements in carbonaceous shales plot off the trend and are enriched from 3 to 10 times relative to the grey shales. These elements were partly introduced with detrital aluminosilicates and partly concentrated by adsorption onto organic matter.

Group 3 – Mo and P (Fig. 11a, b); these trace elements exhibit no correlation with Al in the sandstones and grey siltstones, but show considerable enrichment in the carbonaceous shales. In this case there was no detrital contribution, but only adsorption onto organic matter, when present.

Group 4 – Co and Sr (Fig. 11c); these trace elements exhibit a negative correlation with Al and a positive correlation with Ca, indicating that they are predominantly concentrated in the minor carbonate cement of the sandstones and siltstones.

Some trace elements could not be placed in these groups because: (a) they were not analysed, for example, rare earth elements, (b) they were contaminated during the crushing process, for example, W, or (c) they exhibited no clear trend related to Al content.

Most trace elements, including gold, arsenic and sulphur, show a positive relationship with organic carbon (Fig. 12), which relates to their fixation by both organic matter and pyrite during sedimentation and diagenesis (see below).

Pyrite textures, composition and deportment of trace elements

Diagenetic pyrite in the sedimentary rocks occurs as fine-grained rounded or nodular aggregates (0.5–2 mm across) aligned parallel to bedding (Fig. 13a), commonly consisting of intergrowths of acicular and extremely fine cubic pyrite and marcasite (Thomas et al. 2011). Framboidal pyrite is not common and only occurs in particular shale units (Li et al. 1998).

Hydrothermal pyrite in the sedimentary rocks commonly is coarser grained (0.3–5 mm), euhedral and frequently internally zoned, in places overprinting the metamorphic fabric of the host shale, which may be highlighted by acid etching (Thomas et al. 2011; Fig. 5). In many cases the euhedral pyrites have an anhedral porous core composed of original diagenetic pyrite recrystallized during metamorphism (Fig. 14a), surrounded by euhedral hydrothermal pyrite with growth zoning or an internal metamorphic fabric. Hydrothermal pyrite is more common in sandstone units, presumably reflecting their high effective porosity.

Laser ablation ICPMS mapping and >2000 spot analyses were carried out on diagenetic and metamorphic–hydrothermal pyrites from Bendigo (Large et al. 2009; Thomas et al. 2011) to assist in

Fig. 8. (a) $K_2O$–$Al_2O_3$ for the Bendigo turbidites exhibits a curvilinear trend passing close to GLOSS (global subducting sediment), WAS (Wedepohl’s average shale) and NASC (north American shale composite). (b) The high Rb/K ratio for Bendigo turbidites compared with shale averages (WAS, PASS, GLOSS and NASC) is considered to be characteristic of ancient and highly weathered sources (McLennan et al. 1990). (c) $K_2O$–$Na_2O$ shows a broad inverse trend in the Bendigo turbidites.

Group 2 – V, U, Ni, Zn, Cu, Bi, Pb, S, Se, Ag and Au (Fig. 10); trace elements in sandstones and grey siltstones–shales exhibit a positive linear correlation with Al, as in group 1. However trace elements in carbonaceous shales plot off the trend and are enriched from 3 to 10 times relative to the grey shales. These elements were partly introduced with detrital aluminosilicates and partly concentrated by adsorption onto organic matter.

Group 3 – Mo and P (Fig. 11a, b); these trace elements exhibit no correlation with Al in the sandstones and grey siltstones, but show considerable enrichment in the carbonaceous shales. In this case there was no detrital contribution, but only adsorption onto organic matter, when present.

Group 4 – Co and Sr (Fig. 11c); these trace elements exhibit a negative correlation with Al and a positive correlation with Ca, indicating that they are predominantly concentrated in the minor carbonate cement of the sandstones and siltstones.

Some trace elements could not be placed in these groups because: (a) they were not analysed, for example, rare earth elements, (b) they were contaminated during the crushing process, for example, W, or (c) they exhibited no clear trend related to Al content.

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understanding the diagenetic and hydrothermal history of the deposit. Previous LA-ICPMS studies have shown that diagenetic pyrite in the black carbonaceous shales at Bendigo are enriched in invisible gold and arsenic with mean values of 0.61 ppm Au, 1300 ppm As and Au/Ag >1 (Thomas et al. 2011). Other trace elements enriched in the diagenetic pyrite are Mn, Zn, Mo, Cu, V, Ba, Ag, Cd, Ti, Co, Ni, Bi, Pb and Te. In contrast, euhedral and metamorphic pyrite in the sedimentary rocks commonly overgrows the diagenetic pyrite and contains lower contents of most trace elements, except close to gold–quartz reefs (within 50 m), where the pyrite has thin rims enriched in invisible Au–As (Thomas et al. 2011).

In this study, we have extended the database of LA-ICPMS pyrite compositions for the carbonaceous shales from the distal drill holes NBD005 and NBD186 well removed from the ore deposits, and present new data on the LA-ICPMS trace element composition of the sedimentary rock matrix of the shales (Table 2). Trace element maps of typical pyrite types and associated matrix are also given in Figures 13 and 14. A bar-chart of the

| Table 1. Whole rock and trace element mean values for the three Bendigo turbidite facies |
|-----------------------------------------------|
| **Bendigo Turbidites** | Sandstone  | Grey shale | Black shale |
| Number of analyses wt% | 19 | 47 | 17 |
| SiO₂ | 77.7 | 66 | 65.5 | 58.9 |
| TiO₂ | 0.6 | 0.7 | 0.7 | 0.8 |
| Al₂O₃ | 10.3 | 17.7 | 16.8 | 16.7 |
| Fe₂O₃ | 3.4 | 6.7 | 6.4 | 6.9 |
| MgO | 2.3 | 2.9 | 2.6 | 2.6 |
| CaO | 0.6 | 0.2 | 0.5 | 2.2 |
| Na₂O | 1.4 | 0.7 | 0.5 | 1.6 |
| K₂O | 2.1 | 4.7 | 5.3 | 3.6 |
| TOC | 0.02 | 0.1 | 0.8 | 0.2 |
| Tot S | 0.1 | 0.2 | 0.7 | 0.2 |
| ppm | Ag | 0.08 | 0.14 | 0.24 | 0.07 |
| | As | 9.1 | 21 | 45 | 10 |
| | Au | 1.7 | 4 | 8.9 | 2.3 |
| | Ba | 380 | 932 | 924 | 580 |
| | Bi | 0.2 | 0.7 | 0.8 | 0.1 |
| | Co | 39 | 24 | 28 | 19 |
| | Cr | 37 | 84 | 91 | 90 |
| | Cu | 18 | 43 | 96 | 45 |
| | Mn | 250 | 347 | 304 | 850 |
| | Mo | 0.2 | 0.3 | 2 | 1 |
| | Ni | 23 | 50 | 70 | 68 |
| | Pb | 773 | 550 | 1234 | 710 |
| | Pb | 19 | 31 | 40 | 22 |
| | Rb | 102 | 232 | 254 | 140 |
| | Sb | 2 | 4 | 10 | 1.5 |
| | Se | 1 | 1.3 | 2.5 | 0.5 |
| | Sn | 3 | 5.6 | 5.5 | 6 |
| | Sr | 79 | 63 | 82 | 300 |
| | Te | <0.05 | 0.05 | 1.3 | nd |
| | Tl | 0.6 | 1.36 | 1.43 | 0.68 |
| | U | 3.3 | 3.6 | 6 | 3.7 |
| | V | 48 | 112 | 234 | 130 |
| | W | nd | nd | nd | 1.9 |
| | Zn | 63 | 128 | 360 | 95 |
| | Zr | 124 | 114 | 115 | 160 |
| | V/(V + Ni) | 0.68 | 0.69 | 0.77 | 0.66 |
| | V/Cr | 1.30 | 1.33 | 2.57 | 1.4 |
| | V/Ti | 133 | 259 | 574 | 278 |
| | Ni/Co | 0.59 | 2.08 | 2.50 | 3.58 |
| | V + Mo + Ni + Zn | 134 | 290.3 | 666 | 294 |
ratio of mean elemental values in pyrite over mean elemental values in the background shale matrix is given in Figure 15. These data indicate that the following group of trace elements is strongly concentrated in the diagenetic pyrite relative to the shale matrix, in order of concentration factor: As, Pb, Bi, Sb, Au, Co, Ni, Ag and Se. The trace elements Cu and Mo are moderately concentrated in the pyrite, whereas V, Mn, Sn, Tl, Cr, Sr, U, Zn and W are all preferentially concentrated in the shale matrix.

Preliminary data suggests that different textural forms of sedimentary pyrite show minor differences in the metals concentrated. For example framboidal pyrite (e.g. Fig. 13) has elevated Au, Ag, Te and Mo with a high Ni/Co ratio, whereas pyrite nodules with a fine-grained acicular matted texture after marcasite (e.g. Fig. 14) exhibit elevated Au,
As, Ag, Tl and Te, with no Mo and a lower Ni/Co ratio. These differences probably relate to changes in the redox or pH conditions in the muds between formation of different generations of pyrite (e.g. Schieber 2011). Owing to the nature of the fine intergrowth texture of illite and organic carbon in the carbonaceous shale matrix, it was not possible to confirm by LA-ICPMS which trace elements are associated with organic matter compared with clay minerals in the matrix.

**Chemostratigraphy**

Because the two drill holes sampled for this study cut down through the folded stratigraphy (Fig. 6), it is possible to use the whole-rock analyses to develop a chemostratigraphy for the Bendigo turbidite succession (Figs 16–18). The Railway and Emily Shales exhibit the highest levels of organic carbon, with several sampled intervals over 1 wt%. Higher in the stratigraphy, the Western Shale,
Rowes Shale, top of the Kingsley Unit, the Vital Shale and base of the Herchells Unit exhibit organic carbon values from 0.5 to 1.0 wt% (Fig. 16). There is a general trend of decreasing maximum organic carbon in the shale units, stratigraphically upwards from the Emily Shale to the Herchells Unit. However the shale units below the Emily Shale (Ethan, Owen and Jude Shales) have much lower organic carbon contents of less than 0.2 wt%.

The intervals with higher organic carbon (>0.5 wt%), from the Railway Shale to Herchells Unit, also have significantly elevated levels of V, U, Zn, Cu, Mo, As, Bi, Pb, Ag and Au (Figs 17 & 18), compared with the sandstone units and lower shale units. In Figures 17 and 18, V, Mo and Cu provide the best match with the organic carbon peaks: Zn, U, Mo and Cu give a partial match.

Discussion

Sedimentary Environment

Where deep marine/channel–levee systems have been documented in modern oceans, they commonly occur in fields of vertically stacked, laterally overlapping elements representing sequential channel abandonment (e.g. Wynn et al. 2007). Although our detailed sedimentological data is restricted to the Bendigo Goldfield, its similarity to other areas of Castlemaine Group (e.g. Cas et al. 1988; Ferguson & Vandenberg 2003) suggests that this facies architecture applies throughout the succession.

The recognition that the Bendigo Goldfield is centred on a deep marine channel–levee system has a number of implications for the genesis of the Au deposits.

Firstly, the importance of the black shale units in the Bendigo Goldfield at both a deposit and an outcrop scale has been stressed by Cox et al. (1991a, b). The current explanation for the abundance of black shales in the Castlemaine Group in general is that the area was a site of deep ocean upwelling providing a nutrient-rich environment that promoted high organic productivity (Fergusson & Vandenberg 2003). However, large-scale upwellings in modern oceans are generally located at ocean margins (e.g. Wollast 1998), which is inconsistent with the sedimentological evidence of an abyssal basin plain setting for the Castlemaine Group (e.g. Cas et al. 1988). The existence of channel–levee complexes offers an alternative explanation for the abundance of black shales, because turbidity currents moving down the channels in moderns systems carry organic matter as well as sediment in suspension. For example, a turbidity current event in the Zaire channel in 2001 increased the amount of organic carbon in the bottom waters to 100 times normal levels for tens of kilometres around the channel axis in 4000 m of water (Khripounoff et al. 2003). Much of the organic material in this case was terrestrial in origin, and it has been proposed that its introduction during such events may account for oxygen-depleted...
bottom waters in the region. Although terrestrial organic matter was not available in the Ordovician, marine organic matter sourced from relatively shallow-water high-productivity areas on the continental shelf/slope region could have been incorporated into the terrestrial to deep marine sediment transport pathway. This organic matter would have been carried in suspension along with the clay and mud fraction clastic material, and then settled to form fine-grained hemipelagic deposits in the period following turbidity current events. As illustrated by the contemporary example from the Zaire channel, such a system could be self-sustaining to a degree, in the sense that bottom water anoxia resulting from oxidation of a proportion of organic matter in the water column could have contributed to the preservation of the organic fraction that was incorporated into the bottom sediments.

In contrast to the black mudstone-rich facies association that dominates the Castlemaine Group, Fig. 12. V, Ni, Zn, Cu, Ag and Au each show a positive correlation with organic carbon across the turbidite facies. This is interpreted to indicate that these elements (plus Mo, As, Bi, Pb, Sb and U, not shown) were originally concentrated in the turbidites by adsorption onto clays and organic matter during sedimentation and diagenesis.
that we argue reflects proximity to the channel–
levee complexes that introduced abundant fine-
grained sediments and organic carbon, areas where
chert is the more common fine-grained facies
may have been more distal. This would explain
the dominance of truly pelagic sedimentation of
the remains of marine organisms in these areas
(e.g. Cas et al. 1988), because it would be
less diluted by abundant fine-grained material
periodically introduced via the channel–levee
complexes. Our model is therefore testable, in
that it predicts that the facies association will be
different in these areas where organic rich shales
are less common; for example the deposits that
represent the main channels (i.e. the massive sand-
stone–granulestone, amalgamated sandstone and
thick-bedded sandstone facies) should be less
common.

Fig. 13. LA-ICPMS maps of selected trace elements in sedimentary pyrite from sample NDB186–640 m. (a) Reflected
light image of pyrite (cream) in a shale matrix (grey). The pyrite forms a nodule about 1 mm long, aligned parallel with
the bedding. Pyrite within the nodule has a fine acicular matted texture indicative of precursor marcasite. (b) Trace
element maps in calculated ppm values. Gold, As, Ag, Te, Co and Ni are preferentially concentrated in the pyrite nodule.
Zinc, Mn and V are preferentially enriched in the sedimentary matrix. Thallium occurs in both the pyrite and matrix with
roughly the same order of magnitude. Cobalt and Ni are evenly distributed in the matrix, but at one to three orders
of magnitude less than in the pyrite nodule. The uniform high Ag/Au ratio in this pyrite confirms a sedimentary diagenetic
origin. The lack of Mo and lower Ni/Co ratio indicate a less reduced sedimentary environment than the frambooidal
pyrite in Figure 14.
Fig. 14. LA-ICPMS maps of selected trace elements in sedimentary pyrite from sample NDB005–148 m. (a) Reflected light image of pyrite (cream) in a shale matrix (grey). The core of the pyrite is an aggregate of framboids, and is most likely the primitive texture of sedimentary pyrite deposition. The surrounding euhedral pyrite grew or crystallized during late metamorphism, as indicated by the fabric in the pyrite, which parallels the metamorphic fabric in the sedimentary matrix. (b) Trace element maps in calculated ppm values. Gold, Ag, Mo and Te are preferentially concentrated in the original framboidal core of the pyrite. Arsenic, Co, Ni and Se are preferentially enriched in the euhedral pyrite overgrowth; note the growth zoning in As, Co and Ni compared with Se, and the detectable levels of Ni and Co in the sedimentary matrix. Zinc, Mn, V and Tl are preferentially enriched in the sedimentary matrix, with little or no concentration in the pyrite. Note the high Ag/Au ratio, which is typical of sedimentary pyrite. The elevated Mo and high Ni/Co ratio in the pyrite core suggests growth in a sub-oxic to anoxic environment.
**Palaeo-redox conditions**

A variety of geochemical proxies have been developed to interpret redox conditions of fine-grained sediments during deposition and diagenesis. Palaeo-redox is important in the context of this study as it controls H₂S availability, growth of diagenetic pyrite and concentration of particular trace elements. The most common redox proxies include a combination of C org., degree of pyritization, reactive Fe content, Fe/Al, Mo, V, U, V/Cr, Co/Ni and V/(V + Ni) (e.g. Jones & Manning 1994; Rimmer 2004; Algeo & Maynard 2004; Tribovillard et al. 2006; Meyer & Kump 2008; Algeo & Tribovillard 2009; Algeo & Rowe 2012; Algeo et al. 2012).

Application of these various proxies to the Bendigo shales gives mixed results. For example, the relatively low organic carbon content of the black shales (<2.5 wt%) suggests they formed in an oxic to sub-oxic environment (Algeo & Maynard, 2004), as do the uniformly low Ni/Co ratio (Jones & Manning, 1994). However, the elevated V/(V + Ni) ratio of between 0.70 and 0.86 indicates an anoxic environment (Hatch & Leventhal 1992). Because of these somewhat conflicting interpretations, we have used the method proposed by Tribovillard et al. (2006) to interpret palaeo-redox conditions, of utilizing the full suite of trace elements Cr, U, V, Ni, Co, Cu, Zn and Mo.

The low C org., Mo, V and U contents of the grey shales in the Bendigo succession (Figs 3–5) suggest deposition in an oxic environment. This contrasts with the higher organic carbon in the black shales (0.2–2.0 wt%), relative enrichments in V, U, Mo, Ni, Cu, Cr (Figs 3–5), low Mn content and elevated V/Cr and V/V + Ni ratios. Together these data suggest a sub-oxic to anoxic environment for the black shales. The general lack of framboidal pyrite, except for a few intervals, and predominance of nodular diagenetic pyrite and marcasite in the black shales, suggest that there was commonly no free H₂S in the water column and thus it is very unlikely that euxinic conditions existed at any stage.

Recent work (e.g. Algeo et al. 2012; Scott & Lyons 2012) has emphasized the use of Mo and U values and molybdenum isotopes in refining the palaeoproxies. Even though these two elements are enriched in the black shales at Bendigo (Fig. 17), the levels of enrichment are significantly less than reported for recent marine anoxic and euxinic environments (Scott & Lyons 2012; Tribovillard et al. 2006). This suggests that the black shales at Bendigo may have formed under oxic to sub-oxic rather than anoxic or euxinic conditions. This fits with the common presence of diagenetic marcasite in the shales (now replaced by pyrite), which Schieber (2011) considers to relate to dynamic ocean circulation and reoxygenation of the seafloor muds from oxygen in the overlying sea water. An alternative explanation for the low Mo and U is that draw-down of trace elements depleted the sea water values of Mo and other trace elements during the late Ordovician, as found for the Late Devonian in North America (Algeo & Maynard 2004; Algeo et al. 2012).

In summary, the cyclic pattern of organic carbon and redox sensitive trace elements revealed in the chemostratigraphy shown in Figures 16–18 suggests that the bottom water conditions varied from oxic (deposition of the grey siltstones-shale units) to sub-oxic and possibly temporally anoxic (deposition of black shale units), with intermittent reoxy-}

**Capture of trace elements in shales**

The distinct linear relationships for most trace elements plotted against aluminium (groups 1 and 2 discussed previously and depicted in Figs 6 & 7) indicate that, for the main sandstone, siltstone and grey shale facies of the Bendigo turbidites, the trace elements were delivered to the site of sediment deposition attached to detrital clay particles (e.g. Tribovillard et al. 2006). This process accounts for the background trace elements in the grey sediments deposited under relatively oxic conditions. For the organic carbon bearing black shales, two further processes were involved in the capture and high-level concentration of certain trace elements, including gold and silver. From Figure 10 and Table 1 it is apparent that the elements V, Ni, Zn, Cu, Ag and Au (plus U, As, Bi, Pb and Se not shown in Fig. 10) are significantly enriched in the carbonaceous shales (0.2–2 wt% C org) compared with the grey shales (<0.2 wt % C org). This suggests that these elements were probably absorbed onto organic matter where they ultimately formed organometallic complexes, fixing them into the sub-oxic to anoxic black muds on the basin floor (e.g. Vilor 1983; Kriek 1991; Calvert & Pedersen 1993; Zhang et al. 1995; Wood 1996; Nekrasov 1996; Hu et al. 2000; Algeo & Maynard 2004; Tribovillard et al. 2006; Shpirt et al. 2007; Starostin & Yapaskurt 2007). Most, but not all, of these elements are redox sensitive, and were most probably reduced prior to adsorption and complexation into the organic matter (Wood 1996; Algeo & Maynard 2004).

The second process contributing to the fixation of trace elements in the black shales involves
partitioning of certain elements from organic matter into growing diagenetic pyrite, related to oxidation of organic matter accompanying biogenic reduction of marine sulphate to \( \text{H}_2\text{S} \) (e.g. Algeo & Maynard 2004; Tribovillard et al. 2006; Large et al. 2009). The LA-ICPMS analyses of diagenetic pyrite and organic-bearing matrix in the Bendigo black shales (Table 2) indicate that the trace elements originally adsorbed on to organic matter that were most strongly captured by the pyrite were (in decreasing order) As, Pb, Bi, Sb, Au, Co, Ni, Ag, Se and Cu, whereas elements dominantly retained in the organic-clay-bearing matrix were (in decreasing order) V, Mn, Sn, Ti, Cr, Sr, Zn and W (Table 2; Fig. 15). Some elements were found to be present in both the matrix and the pyrite (e.g. Mo, Cu, Co, Ni).

Source of gold in the black shales

Two possible sources for the gold in the black shales need to be considered: (a) epigenetic hydrothermal addition of gold associated with the emplacement of the gold-bearing quartz reefs; and (b) syngenetic accumulation during sedimentation and diagenesis related to trapping of gold on detrital clays, organic matter and diagenetic pyrite.

The epigenetic hydrothermal origin is supported by the fact that the black shales at Bendigo (e.g. Railway Shale and Rowes Shale) are the immediate host rocks to the gold-bearing saddle reefs (e.g. Cox et al. 1991a, b; Boucher et al. 2008), and thus lateral movement of hydrothermal fluids through the shales may have resulted in minor gold enrichment, even at considerable distances (>hundreds of metres) from the gold reefs. Recent research by Thomas et al. (2011) has also shown that sedimentary rocks within 50 m of gold reefs contain euhedral metamorphic–hydrothermal pyrite with gold–arsenic-rich rims that represent the passage of a Au–As-bearing hydrothermal fluid. These pyrites are best developed in the sandstone and

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**Table 2. Mean LA-ICPMS analyses (ppm) of selected major and trace elements in diagenetic pyrite and black shale matrix; Bendigo black shales**

|                | No. of analyses | Mg   | Al   | Ca   | Ti   | V    | Cr   | Mn   | Fe    | Co   |
|----------------|----------------|------|------|------|------|------|------|------|-------|------|
| Diagenetic pyrite | 70             | 1288.4 | 5554.1 | 202.1 | 1283.6 | 11.5  | 19.5  | 31.9 | 465000.0 | 342.8 |
| Black shale matrix | 72             | 9849.4 | 80770.8 | 428.6 | 2562.6 | 239.3 | 79.8  | 134.5 | 21308.0 | 4.4   |

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**Fig. 15.** Based on 128 LA-ICPMS analyses of sedimentary diagenetic pyrite and 71 LA-ICPMS analyses of sedimentary shale matrix, this graph summarizes the mean element enrichment and depletion of elements in diagenetic pyrite compared with elements in the sedimentary matrix of the Bendigo shales. Bars above the central line indicate degree of element enrichment in diagenetic pyrite, whereas bars below the line indicate degree of enrichment in the clay-rich shale matrix.
siltstone facies, but only occur within 50 m of the gold reefs.

However, notwithstanding the possibility of a hydrothermal origin for the gold in the distal black shales, there are a number of reasons why a syngenetic/diagenetic origin is more likely:

(1) The samples chosen for this study were screened to remove those that showed effects of hydrothermal alteration, and epigenetic Au–As enrichment.
(2) LA-ICPMS analyses (Thomas et al. 2011) have shown that diagenetic pyrite in the black shales at Bendigo contain significant gold, averaging 600 ppb.
(3) Hydrothermal gold mineralization and related pyrite at Bendigo has a Ag/Au ratio less than 1, whereas diagenetic pyrite has an Ag/Au

|   | Ni  | Cu  | Zn  | As  | Se  | Sr  | Mo  | Ag  | Sn  | Sb  | W   | Au  | Tl  | Pb  | Bi  | U   |
|---|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|
|   | 557.5 | 240.6 | 310.9 | 706.5 | 68.8 | 8.1 | 4.3 | 7.8 | 0.5 | 146.0 | 1.5 | 0.38 | 0.3 | 1650.4 | 47.1 | 2.7 |
|   | 25.3 | 23.3 | 225.9 | 3.3 | 3.0 | 43.5 | 1.2 | 0.1 | 4.1 | 1.3 | 2.7 | 0.006 | 1.4 | 4.8 | 0.2 | 5.3 |

Fig. 16. Variation in organic carbon with stratigraphic position. The Emily and Railway shales display the highest levels of organic carbon. Note the general decrease in maximum organic carbon stratigraphically up from the Emily Shale to the Herchules Unit. The shales below Emily have very low organic carbon.
ratio greater than 1 (Thomas et al. 2011). All black shale samples investigated in this study had Ag/Au > 1, confirming the diagenetic origin of the gold.

(4) Gold concentration in the turbidites shows a consistent relationship to sedimentary facies, reflected by its correlation with the content of illite and organic carbon (Figs 7 & 9). The sandstones average 1.5 ppb Au, the grey siltstones and shales average 4 ppb Au and the black shales average 8.9 ppb Au.

(5) The pattern of gold concentration in the shales is very similar to the patterns of other elements (V, U, Ni, Cu, As, Mo, Ag, Zn) that are commonly enriched in organic-bearing shales in anoxic sedimentary environments (Figs 7–9; Rimmer 2004; Algeo & Maynard 2004; Tribovillard 2006). Accepting that the gold in the shales is syngenetic in origin begs the question of the ultimate source of the gold. Many previous researchers have suggested that elevated gold (and platinum group elements) concentrations in metalliferous black shales, such as those in mid-continent of USA (Coveney & Martin 1983), the Carlin district USA.

Fig. 17. Variation in trace elements V, Mo, Zn, U and Cu in sandstones and shales with stratigraphic position. Consistently high values of all five trace elements occur in the Emily and Railway Shales in samples with elevated organic carbon content. Less consistent patterns of enrichment are present in Western Shale, base of Rowes Shale, Vital Shale and Herchules Unit.
(Emsbo et al. 1999, 2000) and in southern China (Coveney 2000), are due to hydrothermal exhalation into the sedimentary basin. Others, such as Holland (1979), Mao et al. (2002) and Lehmann et al. (2007), have argued that hydrothermal exhalation is not necessary, but adsorption from sea water onto clays and organic matter is sufficient to account for the metal enrichment. The high Rb/K ratios in the shales, and complete lack of volcanic units, volcanic detritus or syn-sedimentary feeder faults at Bendigo, suggest that hydrothermal exhalation is an unlikely source for the gold in the shales.

On the other hand, the correlation between gold, organic matter and redox-sensitive trace elements in the Bendigo shales supports the adsorption from sea water model.

We contend that the sedimentary environment of the Ordovician turbidites in central Victoria is the key to gold enrichment in the shales. Our studies, coupled with those of Vandenberg et al. (2000) and Boucher et al. (2008), indicate that the upward-fining sandstone-dominated cycles in the Bendigo host succession represent deep-water channel facies analogous to the distal parts of the deep
water sinuous canyon–channel–levee systems described outboard of major river systems from contemporary basin settings (e.g. the Amazon, Mississippi and Congo). These supply huge volumes of mainly fine-grained sediment and organic material via shelf margin canyons, with the fine-grained organic-bearing mudstone facies deposited either in over-bank levees or filling abandoned channels. Gold may be transported in from the eroding hinterland, either adsorbed on detrital clays, or as colloidal particles (e.g. Reznik & Fedoronchuk 2000). These authors propose that turbidity flows may transport the gold from its source, along palaeo-valleys and submarine canyons for over a hundred kilometres, over the continental slope and into deep-water environments.

The thick channel sandstone facies at Bendigo are testament to a high-volume sediment transport system from the source hinterland, whereas the organic-bearing levee bank and channel fill shale facies represent the ultimate sink for gold, arsenic and other redox-sensitive trace elements transported within the Ordovician canyon–channel–levee system (Fig. 4). However gold accumulation in the black shale facies never reaches ore grade. A later up-grading mechanism such as greenschist facies deformation or granite intrusion is required to liberate the gold from the shales and concentrate it in favourable structural sites by hydrothermal processes (Large et al. 2011). Starostin & Yapaskurt (2007) propose a very similar multistage model for the concentration of gold in Late Carboniferous to Early Permian black shales from the Verkhoyansk–Kolyma fold belt in western Siberia.

**Conclusions**

Analysis of the stratigraphy and sedimentology of the thick Ordovician turbidite succession in the Bendigo district of Victoria suggests that it records a sinuous channel–levee system developed in an abyssal plain environment. By analogy with modern systems, it would have been the ultimate product of a giant river–canyon–fan system sourced from the Delamarian highlands to the west. Gold eroded from this hinterland was transported attached to detrital clays, and deposited in the over-bank and channel fill mudstones by reduction, adsorption and complexation with organic matter. The resultant black shale facies at Bendigo varies from 0.2 to 2 wt% organic carbon, has a mean gold content of 8.9 ppb, and is enriched in other redox-sensitive trace elements, including V, As, Bi, Mo, Ni, Ag, Zn, Cu, Sb and U. The anomalous syngenetic Au and As content of the black shale facies at Bendigo makes the Ordovician sedimentary succession in Central Victoria a viable source for the Au and As in the orogenic quartz vein gold deposits.

This research was undertaken with financial support from the Australian Research Council and Bendigo Mining. Thanks to H. Thomas for assistance with drill core sampling and L. Danyushhevsky and S. Gilbert for providing technical advice on LA-ICPMS analyses. Thanks also to four reviewers who provided critical and useful comments to assist in improving the manuscript.

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