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Hydrocarbon-derived imprints in olistostromes of the Early Serravallian Marnoso-arenacea Formation, Romagna Apennines (northern Italy)

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Abstract Evidence of hydrocarbon venting within slumped bodies associated with the siliciclastic, dominantly turbiditic, Marnoso-arenacea Formation (Umbria-Romagna structural domain, Romagna Apennine, northern Italy) is documented with sedimentological, faunal, and geochemical data. Specifically, ^{13}C-depleted carbonate concretions and limestones and clusters of chemosynthetic clams (Vesicomyidae) have been identified in the marls of the Le Caselle Olistostrome and other slumped bodies contained within the Early Serravallian section of the Marnoso-arenacea Fm. Most of the olistostrome marls and limestones are extrabasinal and must have slid from a source area located several kilometers southwest of their present position. Thus, they presumably pertain to the Vicchio Marls Formation of the northeastern (outer) Tuscan structural domain, with possible minor contributions from the epi-Ligurian Bismantova Fm. It is suggested that venting of methane in the source area of the olistostromes permitted the establishment of exotic chemosynthetic communities and promoted the precipitation of carbonate concretions and limestones. According to the field evidence, these materials were later subjected to multistep downslope remobilization and were eventually carried into the Marnoso-arenacea basin through gravity mass transport.

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Introduction

The importance of methane cold-venting in the maintenance of specialized chemosynthetic benthic communities and the formation of distinct sedimentary structures and carbonate buildups has been recognized since the discovery of such occurrences at the base of the Florida escarpment in the mid-1980s (Paull et al. 1984). Since then, similar discoveries in other parts of the modern oceans have prompted scientists to recognize their fossil counterparts in the geologic record (e.g., Beauchamp and von Bitter 1992).

The objective of this study is to present new data on a site located in the Romagna Apennines of northern Italy, about 40 km southwest of Forlì (Fig. 1A, inset map). The proposed fossil clues to paleoventing do not belong to the already recognized sequence of methanogenic *Lucina* limestones so widespread in the Miocene of the Italian peninsula (Ricci Lucchi and Vai 1994), further demonstrating the importance and diversity of deep-sea methane venting in the Miocene of the Apennines.

Geological setting

The Le Caselle Olistostrome (LCO) is included within the Marnoso-arenacea Fm, in a stratigraphic position about 1070 m above the lithostratigraphic marker referred to as Contessa in the literature (Ricci Lucchi 1975). Both the LCO and the enclosing sequence are Early Serravallian in age, as indicated by biostratigraphic dating (Berti and Cuzzani unpublished data).

From a tectonic point of view, the mostly vertical position of the exposed section of the LCO (striking N110E) is related to the backlimb of a tight footwall syncline underlying a main thrust front (Fig. 1A and B). The syncline is one of best examples of the typical fault-propagation folds known as Romagna Structure ["struttura romagnola," according to Signorini (1940)]. This tectonic style, regularly

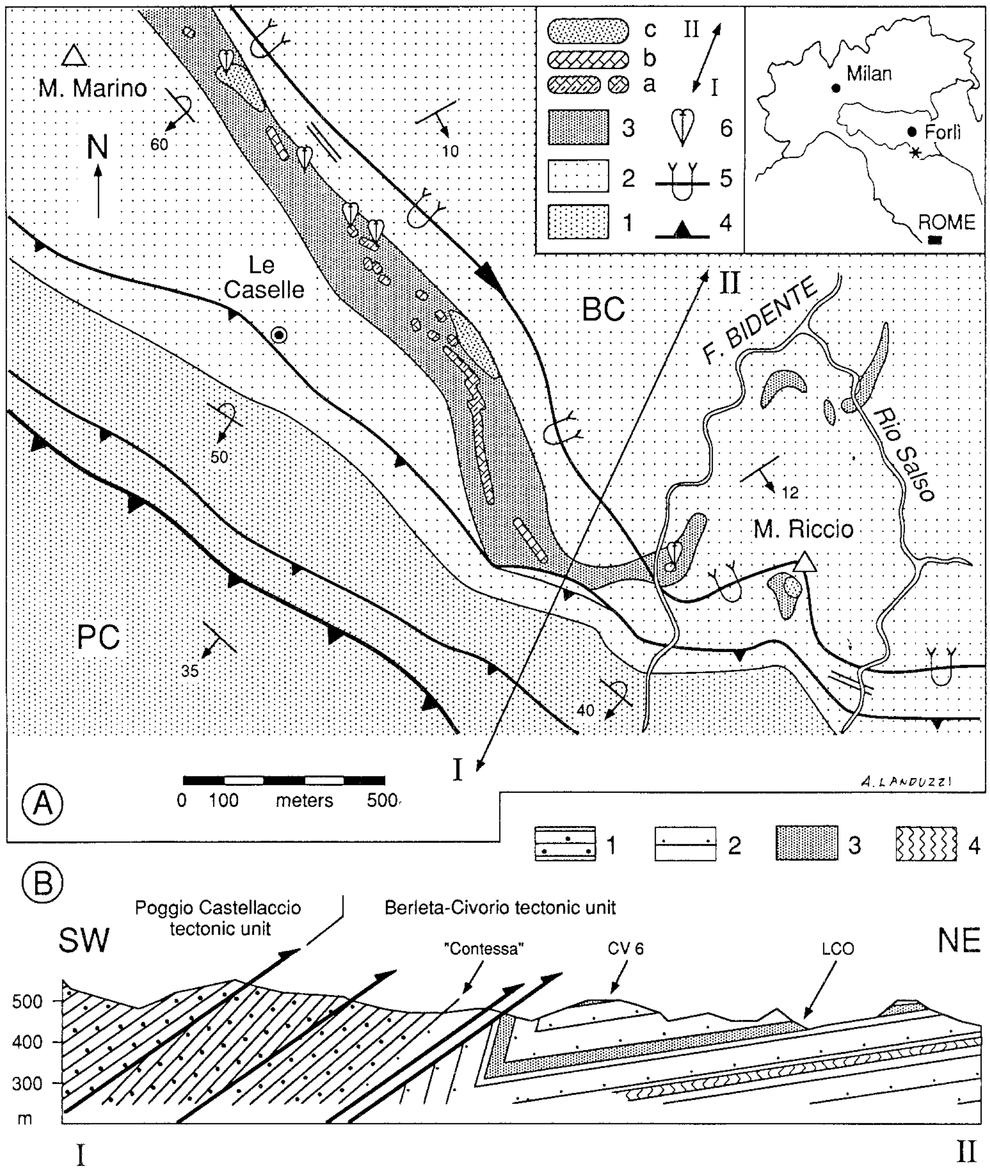


Fig. 1 Schematic geologic map and cross section of the study area. **A** Geologic map: 1 = Marnoso-arenacea Fm, pre-Contessa unit; 2 = Marnoso-arenacea Fm, post-Contessa unit; 3 = Le Caselle olistostrome (extrabasinal marls); a = Olistolith A; b = Olistolith B; c = Olistolith C; 4 = thrust fault; 5 = recumbent syncline; 6 = fossil bivalves; I–II = cross-section track. PC = Poggio Castellaccio tectonic unit; BC = Berleta-Civorio tectonic unit. **B** Cross-section: 1 = Marnoso-arenacea Fm, pre-Contessa unit; 2 = Marnoso-arenacea Fm, post-Contessa unit; 3 = Le Caselle (LCO) and CV 6 olistostromes; 4 = intraformational slump

verging to the northeast, has been established in Late Messinian by thrusts detached on a common level close to the base of the Marnoso-arenacea Fm. Younger and more deeply detached compressional deformations have superimposed on the Late Messinian structure in Early Pliocene, Middle Pliocene and Early Pleistocene. Subsequently, the compressional style has been replaced by an extensional one, which is still active today (Vai 1988; Landuzzi 1991; Capozzi et al. 1992).

The Late Messinian compressional style typifies the Umbria-Romagna structural domain as an imbricate fan of several tectonic units (Van Wamel and Zwart 1990; Capozzi et al. 1992). In particular, the studied LCO outcrops pertain to the Berleta-Civorio unit of southeastern Romagna, which has been thrusted over by the more southwestern (inner) Poggio Castellaccio unit (Fig. 1A and B). Further to the southwest, the Umbria-Romagna structural domain has been thrusted over by the northeastern (outer) Tuscan structural domain, which has in turn been thrusted over by the Ligurid allochton. Although the LCO is actually enclosed in the inner Umbria-Romagna domain, its lithofacies bearing evidence of methane ventings are all extrabasinal and have certainly been displaced by gravity transport from the southwest. Thus, the Miocene sequences pertaining to the outer Tuscan domain (Vicchio Marls Fm) and the semiallochtonous sequences deposited on the Ligurid allochton (epi-Ligurian units, as the Bismantova Fm) are both relevant to the LCO provenance. As a consequence, the synsedimentary tectonic pulses that affected the outer Tuscan domain are important clues to the paleoenvironmental setting of the LCO source area, as well as the eventual block-sliding of the LCO into the Marnoso-arenacea basin.

The outer Tuscan domain has been primarily displaced during the Burdigalian compressional phase of the Apennine orogeny, when the foredeep axis was shifted northeastward from the Tuscan Cervarola-Falterona turbidite basin to the Marnoso-arenacea basin (Ricci Lucchi 1975, 1986). In the field, this Burdigalian phase is marked by a clear hiatus in deposition within the Burdigalian-Early

Serravallian Vicchio Marls Formation, which has been interpreted as a closure facies over the Cervarola-Falterona clastic wedge (Pizziolo and Ricci Lucchi 1992). A minor regional peak of compressional activity, the upper Early Serravallian event, is documented by synsedimentary thrusting of both the outer Tuscan domain and the Ligurid allochton over the Marnoso-arenacea Fm of southwestern Romagna. Indirect field evidence supporting the same tectonic event comes from the emplacement of Ligurian and Tuscan olistostromes, including LCO, over the whole inner part of the Marnoso-arenacea basin.

Olistostromes: Stratigraphy and lithofacies

Le Caselle olistostrome

The exposed section of the LCO, about 150 m thick, is remarkably uniform over a distance of about 1.5 km (Fig. 1A). Further to the northwest, the olistostrome has been eroded away because of the axial culmination of the encompassing syncline. Further to the southeast, the olistostrome has been progressively involved in thrusting and consequent tectonic burial by the Poggio Castellaccio unit. Following the inferred southwest–northeast sliding direction for 700 m, the olistostrome thickness gradually decreases to 5 m.

The LCO consists primarily of marls enclosing dominant carbonate and subordinate siliciclastic olistoliths. An unusual feature of the LCO and the stratigraphically younger olistostromes is represented by the occurrence of limestone concretions (see below) and pebbles, up to some tens of centimeters in diameter. In some cases, the pebbles are broken and reworked; otherwise, they occur as irregularly rounded clasts, composed of biomicrite with variable amounts of siliciclastic (and subordinate calcareous) silt and sand. These clasts can be found either scattered within the marly matrix or reworked into bedded horizons emplaced by gravity mass flows (Figs. 2 and 3). A few hybrid calcarenite pebbles contain abundant bioclasts of probable shelf provenance (algae, large foraminifera, bryozoans and mollusks). The fabric of most pebbles reveals a thin, parallel lamination, whereas the upper surfaces are often pitted and reddish, suggesting exposure on the sea bottom (hardgrounds?) and the possible action of corrosive waters.

In spite of evident dismembering and deformations due to submarine sliding processes, the internal setting of the LCO is not completely chaotic. Indeed, the average bedding position is well defined, and the vertical sequence of lithofacies remains roughly the same throughout the exposed section of the olistostrome (Fig. 1A). This makes it possible to interpret the LCO as a stratigraphic sequence of three major olistoliths (in the order A-B-C, as indicated below: see scheme in Fig. 2). These olistoliths are inferred to have emplaced as separate events in the source area of the LCO. Subsequently, when the LCO slid as one unit into the Marnoso-arenacea basin, the stratigraphic order of the olistoliths is likely to have been preserved.

Olistolith A

Olistolith A is composed of a 80-cm-thick, graded arenite–siltite bed of hybrid composition (rich in micrite), sharply separated from an underlying 20-cm-thick marly limestone bed. The arenite–siltite bed shows a probable incomplete Bouma sequence, with ripple-scale cross-bedded lamination (Tc interval) and thin flat-bedded lamination (Td interval). In addition, the base of the same bed is locally marked by a *Palaeodyction* ichnofacies (Cuzzani 1992). All these data suggest a deep-sea environment fed by sandy turbidity currents, whereas the marly limestone can be interpreted as a turbiditic fine tail and/or as hemipelagite.

Microscopic observations of arenite samples clearly show that they are mostly composed of sandy biomicrites with siliciclasts and fecal pellets embedded in cryptocrystalline micrite; pyrite framboids are common and ubiquitous. In many places, Olistolith A consists of monomictic breccias related to sedimentary dikes and extensional fractures, showing no preferential orientation with respect to bedding planes. Breccia clasts are composed of either hybrid arenite or marl. Microscopic observations of the cement reveal the presence of light brown prismatic spar crystals (0.5–1 mm long), grown orthogonally to the clast boundaries. This first-generation calcite growth must have occurred in fluid-filled cracks, possibly resulting from violent outbursts of methane-bearing fluids breaking up the upper crust of already lithified sediments (autoclastic–breccia processes; Hovland 1987). On the other hand, some polymictic breccias occur locally, which can be the result of displacement by synsedimentary slides of the previously described lithologies.

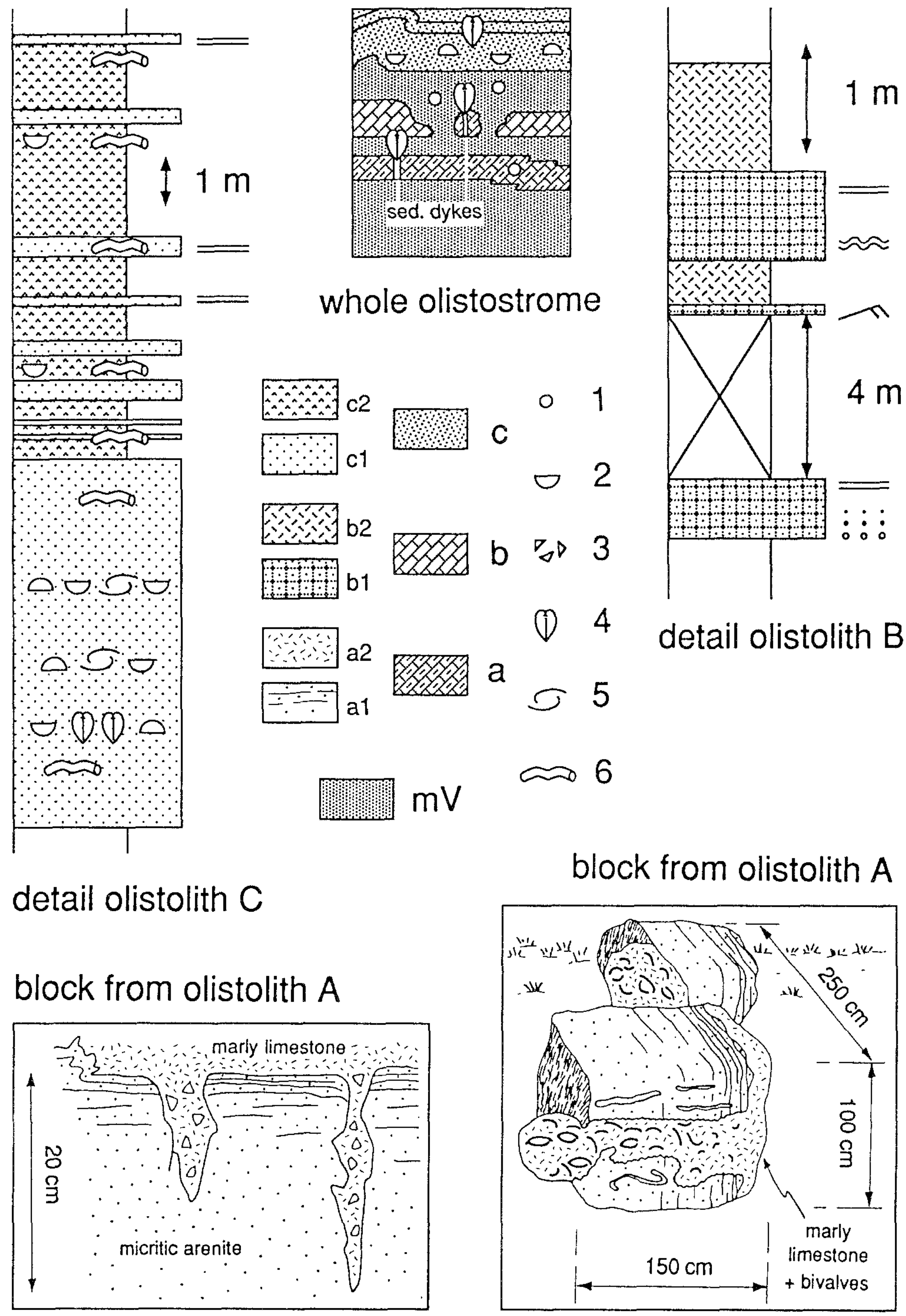
One of the most important findings within the arenite–siltite bed is the recurrence of subspherical nodules (1.5 cm in diameter) consisting of radially arranged calcite and aragonite crystals. Similar nodules have been observed in presentday cold-vent environments where, in response to aerobic methane oxidation, aragonite occurs both as pore-filling cement and as botryoids (Roberts et al. 1990). The pyrite occurrences within the arenite–siltite bed may be related to alternating anoxic and oxic phases, marked by erosion as well as cracking-and-filling of sedimentary crusts (Hovland 1987).

Large bivalves, mostly mussels (?*Bathymodiolus* sp.) and clams (?*Calyptogena* sp.) have been found within an isolated block belonging to Olistolith A (Fig. 2). In particular, such faunas are scattered within the marly limestone filling of a tens of centimeters wide sedimentary dike crossing the arenite–siltite bed. This occurrence further supports the methane cold-venting hypothesis (Taviani et al. 1992) for the turbidite facies assumed as source area of Olistolith A.

Olistolith B

Olistolith B is composed of interlayered turbiditic couples of light brown calcarenite and light gray marl, and reaches an overall thickness of about 5 m. The calcarenite consists of limestone lithics and biogenic debris, embedded within a micrite matrix.

Fig. 2 Stratigraphic relationships between olistoliths within Le Caselle Olistostrome. mV = light-colored silty marls with slump-deformed thin-bedded turbidites; a = whole Olistolith A; **a1** = partly laminated, disturbed hybrid arenite–siltite (turbidite); a2 = yellowish, homogeneous marly limestone; b = whole Olistolith B; b1 = well-bedded calcarenites (turbidites); b2 = light-colored limy marls; c = whole Olistolith C; **c1** = poorly cemented sands (grain flows); c2 = blue-gray silty marls; 1 = limestone concretions; 2 = limestone and hybrid clasts (pebbles); 3 = limestone breccia; 4 = chemosynthetic clams; 5 = reworked shallow water faunas; 6 = bioturbations



Olistolith C

Olistolith C comprises coarse, poorly cemented sandstones and silty marls, attaining an overall thickness of about 14 m (Fig. 2). As a whole, this sequence can be interpreted as an interlayering of grain flows and thin turbidites, revealing a proximal depositional environment (Berti and Cuzzani unpublished data).

Within the thickest sandstone bed (about 6 m thick), limestone and hybrid pebbles are confined to three separate parallel-bedded horizons (Fig. 3), marked by concentrations of millimeter-sized degraded pyrite nodules. Pebbly horizons contain rare centimeter-sized whitish angular blocks, made of the same marly limestone present also in Olistoliths B and A. The thickest sandstone bed is also typified by a fossil concentration of vesicomyid clams (see below).

Other olistostromes

The LCO is the thickest of a series of submarine slumps that are enclosed in the post-Contessa Serravallian

Fig. 3 Pebbly horizons within Olistolith C

Marnoso-arenacea Formation. Indeed, nine of these bodies occur up-section in a 150-m-thick envelop (starting from the LCO), and six of them enclose an extrabasinal component similar to the material of the LCO itself (Berti and Cuzzani unpublished data). These six olistostromes contain carbonate pebbles, and sparse limestone breccia blocks similar to those described in the section on Olistolith A. The most prominent of the six is the olistostrome CV 6 (110 m up-section from the LCO; see Fig. 1B) that is composed of limestone breccia and contains a great number of lucinid bivalves.

Discussion

Faunal evidence supporting methane paleo-venting

The presence of synsedimentary methane cold-venting in the Early Serravallian source area of the Le Caselle olistostrome is supported by the occurrence of a monospecific assemblage of clams belonging to the family Vesicomyidae (Taviani et al. 1992; Aharon et al. 1993) within the thickest grain flow of Olistolith C. The clams form a bed about 20 cm thick and are mostly represented by articulated adult shells of roughly similar size. We counted 12 shells over an horizontal distance of 2 m. Some of the shells have their commissural plane parallel to the bedding and only a few are in a vertical (growth) position. We were unable to retrieve complete shells from the outcrop to carry out a full morphological study. Shells were intimately embedded within concretions of carbonate-cemented silty sand, and attempts to remove whole specimens led to breakage of the fossils. In particular, the hinge characteristics of the vesicomyids, an important diagnostic feature at specific and supraspecific levels (Horikoshi 1989), are still unknown, thus impeding a generic identification of the fossil material. However, external shell features fit well those of *Calyptogena l.s.* The general outline closely resembles species of the Pacific *Calyptogena* described from Japan (Okutani and Egawa 1985; Okutani and Metivier 1986).

Vesicomyids are highly specialized deep-sea bivalves harboring chemosynthetic bacteria in their gills (e.g., Boulegue et al. 1987) and are distributed at hydrothermal sites, cold vents, and anoxic deep-sea sediment areas in the Pacific and Atlantic oceans (e.g., Turner 1985; Gage and Tyler 1991). Stable-carbon isotope studies of two Pacific species (*Calyptogena phaseoliformis* and *Calyptogena soyoae*) document their symbiosis with methylotrophic bacteria acting as primary producers (Saino and Ohta 1989). Nutrition of these clams depends, therefore, upon chemosynthetic processes involving oxidation of hydrogen sulfide and/or methane, a view supported also by histological and ultrastructural work performed on *C. laubieri* and *C. phaseoliformis* (Fiala-Medoni and Le Pennec 1989).

Direct observations of living vesicomyid assemblages have shown that in fine-grained sediments surrouding cold-venting areas, a setting similar to the Miocene source area of the LCO, these bivalves are adapted to a very shallow epifaunal life (e.g., Okutani and Egawa 1985; Hashimoto et al. 1989). In situ observations of *C. soyoae* have shown that adults of this species live half-buried in the sediment is a vertical posture (Okutani and Egawa 1985; Hashimoto et al. 1989), while juveniles are completely embedded in the sediment. In some occasions, vesicomyids have been observed crawling over the sea bottom and actively plowing the sediment surface (Rosman et al. 1987).

Vesicomyids are very gregarious animals, often forming dense monospecific communities. Photographic and video records from cold vents on the Louisiana–Texas continental slope (Rosman et al. 1987) and from the subduction zone off Japan (e.g., Hashimoto et al. 1989) have documented the tendency of *Calyptogena* to aggregate in dense colonies. Fossil vesicomyid-bearing (*Akebiconcha kawamurai*) shell beds from the Pliocene of Japan show remarkable analogies with the presentday vesicomyid assemblages, consisting of an equal high density of clams and a strong dominance of one species with respect to the total macrofauna (Nobuhara and Tanaka 1993). The *Calyptogena* bed within the LCO is no exception to the general spatial organization of cold-vent-related vesicomyid assemblages. In fact, the clams consist of a distinctly monospecific assemblage with apparently no other macrofaunal element mixed with them. This observation may reflect the effect of postmortem selective diagenesis, although it is conceivable that originally the vent community included only a few taxa. By analogy with modern cold-vent communities, we cannot rule out the possibility that other taxa (e.g., vestimentiferans, brachyurans) coexisted with ?*Calyptogena* sp., but it is likely they left no fossil trace because of the absence of hard parts.

From the paleontological point of view, there are no indications of significant postmortem transport. Thus, taking into account the sedimentological evidence, it is likely that the faunal assemblages were incorporated into the thickest grain flow of Olistolith C as large soft-sediment slabs. At present, vesicomyid clams occupy a wide bathymetric range between 100 and 6000 m, with most species distributed between 1000 and 3000 m (Boss and Turner 1980). Under these circumstances it is not possible to state the depth at which the fossil ?*Calytogena* assemblage thrived. We postulate a considerably deep setting, perhaps in the order of 1000 m or more.

The ?*Calyptogena* assemblage may represent an ephemeral event, as inferred from the absence of similar shell beds in the rest of Olistolith C. If so, conditions favorable to the development and maintenance of the cold-vent community in the specific source area of the Le Caselle Olistostrome should have lasted only months, or perhaps a few years at the most. However, cold venting able to sustain specialized chemoautotrophic communities was a common phenomena in other parts of the basin, as indicated by the wide spatial and stratigraphic distribution of cold-vent faunal beds in the Apennines (Taviani 1994; Ricci Lucchi and Vai 1994).

Geochemical evidence supporting methane paleoventing

Some carbonate samples from the vesicomyid unit (Olistolith C) and the limestone facies of Olistolith A have been analyzed for their stable carbon isotope compositions; the results are listed in Table 1 (also see Fig. 2). The well-preserved shell of ?*Calyptogena* yielded a $\delta^{13}C$ value of $-0.2‰$, which is not particularly depleted in ^{13}C. This result is not surprising because, in contrast to the soft parts that reflect the strongly negative methane $\delta^{13}C$ signature (e.g., Saino and Ohta 1989; Rio et al. 1992), the $\delta^{13}C$ composition of cold-vent bivalve shells does not show a similar depletion in ^{13}C. $\delta^{13}C$ shell values as low as $-8‰$ have so far been observed only in cold-vent mussels (Paull et al. 1984), but never in *Calyptogena* (Kulm et al. 1986; Saino and Ohta 1989). On the other hand, the observed ^{13}C depletions of the concretioned calcite matrix enveloping the vesicomyid shells ($\delta^{13}C = -14.3‰$), and the limestone facies of Olistolith A ($\delta^{13}C = -31.6$ to $-34.9‰$), can be interpreted as an indication of methane venting imprinted on the carbonates of the LCO source area (e.g., Aharon et al. 1992).

Table 1 Carbon isotope compositions of carbonates from the ?*Calyptogena*-bearing unit, Olistolith C and limestone facies of Olistolith A (see Fig. 2)

Sample	Olistolith	$\delta^{13}C$ (‰ PDB)
Shell of ?*Calyptogena*	C	−0.2
Concretioned calcite matrix	C	−14.3
Concretionary clast	C	−6.1
Brecciated micritic limestone	A	−31.6
Infill of ?*Calyptogena* molds	A	−34.9
Laminated calcarenite	A	−31.9
Marly limestone	A	−33.1

Anomalously low $\delta^{13}C$ values of concretions have been interpreted as indications of a possible methanogenic derivation of micrites (Coleman et al. 1981). As already discussed, most concretions and pebbles preserve an internal lamination. The intrusion of methane bearing fluids into semiconsolidated deposits, without deformation of their delicate sedimentary structures, should have been facilitated by a microdiffusion process (Roberts et al. 1990). In this way, methane oxidation should have occurred in conjunction with sulfate reduction in the anaerobic zone, a fact which might explain the abundance of framboidal pyrite observed in thin sections (Hovland 1987).

From a stratigraphical point of view, only a few concretions seem to have preserved their primary relationships to the embedding lithofacies, (refer to mV marls and Olistolith A in Fig. 2). On the other hand, many pebble biomicrites are also interpreted as methanogenic concretions, which have been reworked downslope and mixed with hybrid calcarenite clasts of presumed shallower water origin.

Paleoenvironmental interpretation of Le Caselle olistostrome

The proposed source area of the Le Caselle Olistostrome is identified with the outer Tuscan domain (Fig. 4), located between the deep turbidite Marnoso-arenacea basin to the northeast (Umbria-Romagna domain) and the hybrid carbonate Bismantova Fm shelf to the southwest (epi-Ligurian semi-allochton). In particular, marls and calcarenites within the LCO have no affinities to the enclosing Marnoso-arenacea Formation, whereas they clearly resemble lithotypes from the coeval, Early Serravallian section of the Vicchio Marls Formation, cropping out 15 km southwest of the present study area (Pizziolo and Ricci Lucchi 1992). However, in post-Serravallian times the Vicchio Marls Fm has been eroded away from most of the inferred olistostrome source area, so that today no geometric relationships between Vicchio Marls and LCO can be directly assessed in the field.

A depositional model of the Le Caselle olistostrome, illustrated in Fig. 4, has been derived on the basis of stratigraphic evidence, faunal assemblage, and carbon isotope data discussed in the previous sections (also see Ricci Lucchi and Vai 1994). This model, which is framed in the time lapse between the Burdigalian and the upper Early Serravallian compressional pulses, includes the following scenario:

1. Physiographic conditions affecting primary sedimentation of the oldest LCO lithofacies are likely to have derived from residual compressional tectonics postdating the Burdigalian phase. This is because vast subsiding turbidite environments developed in the Langhian and lower Early Serravallian Vicchio Marls Fm. Methane vents in this turbidite realm are suggested by carbonate concretions and limestone facies within Olistolith A (Fig. 2).

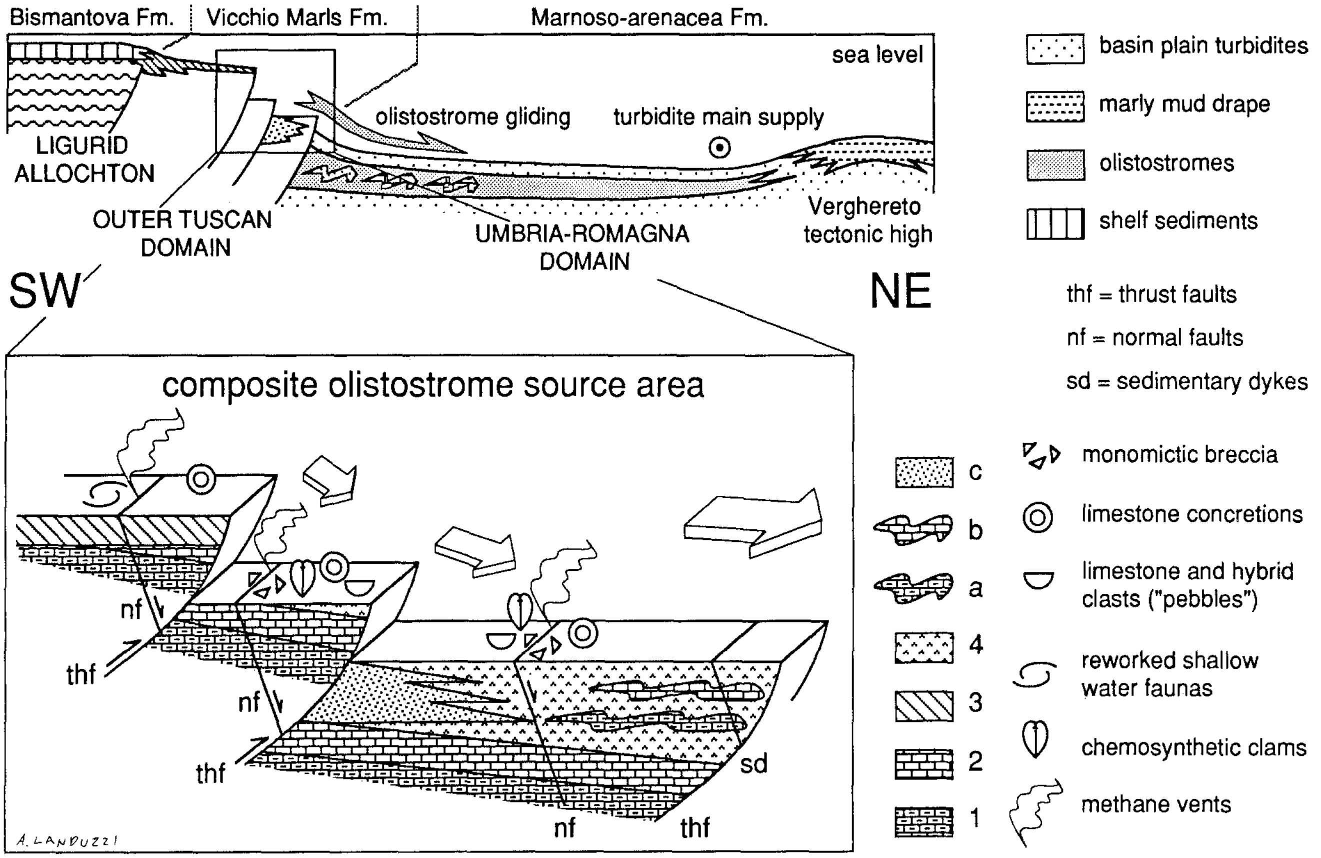


Fig. 4 Schematic paleoenvironmental interpretation of olistostromes. 1 = hybrid turbidites and marly limestones (source for Olistolith A); 2 = calcareous turbidites (source for Olistolith B); 3 = outer shelf—inner slope sediments; 4 = intermediate and outer slope marls; a = Olistolith A; b = Olistolith B; c = grain flows and Olistolith C

2. In the Early Serravallian, isostatic rebound and perhaps extensional tectonics have led to differential uplift, nucleating the articulated depth zones of the composite LCO source area (Fig. 4). Consequently, Vicchio Marls turbidites have been progressively taken over by slope and eventually outer shelf sediments. Tectonic-related evidence of methane vents is provided by the occurrence of mussels and vesicomyids in a sedimentary dike crossing Olistolith A (Fig. 2).

3. Shortly before the upper Early Serravallian compressional event, an increasing gravitational collapse of the outer shelf, as well as the inner and intermediate slope, has been marked by grain-flow and olistolith accumulation within the outer slope marly deposits. Olistolith A (hybrid arenite–siltite) and Olistolith B (calcarenite) are likely to have originated from the previously quoted lower Early Serravallian turbidite sequences, exposed to sliding in the intermediate depth zone of the LCO source area (Fig. 4). The younger Olistolith C is supposed to have derived from penecontemporaneous grain flows in the most proximal part of the outer slope. In particular, the layers of mixed hybrid and limestone pebbles (Fig. 2) within Olistolith C suggest a multisource provenance of the grain flows, from different depth zones subject to methane vents. For instance, terrigenous debris and bioclasts of probable shallow marine origin, occurring within the hybrid pebbles, seem to imply contributions from the outer shelf. Another possible source area, quite possibly subject to vents and perhaps deeper than 1000 m (Fig. 4: intermediate slope?), is indicated by the fossil ?*Calyptogena* sp. assemblage, found in a sandy-silty slab within the thickest grain flow of Olistolith C. The most recent methane venting episode in the lower-slope depth zone of the proposed model (Fig. 4) is indicated by likely non-reworked limestone concretions in the marl unit between Olistoliths B and C (Fig. 2).

4. During the upper Early Serravallian tectonic event, slope marls and enclosed olistoliths slid as one unit (the final Le Caselle olistostrome, Fig. 4; see also Fig. 1A and B) from the lower depth zone of the Vicchio Marls Fm into the Marnoso-arenacea basin. Around 20 km northeast from the proposed source area (Fig. 4), LCO and all younger olistostromes pinched out onto the tectonically active Verghereto-S. Sofia intrabasinal high (Berti and Cuzzani unpublished data), interlayering with some marly slumps coming in the opposite direction.

Conclusions

Sedimentary (autobreccias, concretions), faunal (vesicomyid clams), and geochemical (negative $\delta^{13}C$) evidence support the existence of active methane cold venting in the

Early Serravallian section of the outer Tuscan sequence, occurring within a prevalently deep siliciclastic environment. Preliminary analysis of limestone–pebble microfacies suggests that methane venting has occurred in three bathymetrically distinct depositional environments:

1. A relatively shallow area (outer shelf?), probably located within the outer Bismantova Fm. After their formation, the concretions evolved into pebbles through submarine transport processes and have either been enclosed in grain-flow deposits (Olistolith C) or dispersed in the overall marly matrix.

2. An intermediate depth area (1000 m or deeper) indicated by the chemoautotrophic bivalve assemblages (mussels and vesicomyid clams). Clams have been found as molds in a sedimentary dike of Olistolith A and as clusters in sediment slabs within the grain flows of Olistolith C.

3a. An older deep area in the turbidite realm, indicated by autoclastic breccias, non-reworked concretions, and anomalously negative $\delta^{13}C$ compositions of limestone facies from Olistolith A.

3b. A younger deep area in the distal zone of the grain-flow realm, suggested by non-reworked concretions embedded in marls.

The exact source of the venting methane is uncertain and needs further assessment. As a speculative hypothesis, we suggest a deep-seated methane source, possibly related to hydrocarbon reservoirs within the Mesozoic suite. Under these circumstances, we also suggest that fault and fracture sets, inherited from compressional tectonics or produced by a temporary extensional regime, have acted as conduits for methane venting in the submarine environment.

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