Deposition of megabeds (megaturbidites) and sea-level change in a proximal part of the Eocene-Miocene flysch of central Dalmatia (Croatia)

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ABSTRACT

Some carbonate megabeds of the proximal part of the Dalmatian Eocene-Miocene flysch basin (eastern Adriatic Sea) are underlain and overlain by conglomeratic fan deltas, whereas thinner-bedded turbidites interfinger with fan-delta conglomerates. Fan deltas and interfingering turbidites are interpreted as having been deposited during lowstand(s) and early sea-level rise(s), whereas deposition of megabeds is attributed to periods of accelerated sea-level rises that aborted growth of fan deltas. Attribution of megabeds to periods of sea-level rise is based on the following observations: (1) they overlie some conglomeratic fan deltas, (2) skeletal and lithic debris in megabeds originated by subaerial weathering of older sediments, and (3) their large thickness indicates considerable primary accumulation of debris and low frequency of flows. Sea-level changes inferred from megabeds are different from those related to thinner classical turbidites, and distinction of the two is important for sequence stratigraphy in deep-marine sediments.

INTRODUCTION

Megabeds (megaturbidites, olistostromes) are common in many flysch basins. The role of megabeds in basin filling may be very significant; in the Paleocene-Eocene of the Friuli (Italy), they make up nearly one-half (Tunis and Venturini, 1992), and in the Eocene-Miocene central Dalmatian basin (Fig. 1), they account for one-third, of the total succession (Marjanac, 1989, 1991). There is no generally accepted precise definition of megabeds, but the consensus is that they are defined by exceptional thickness (e.g., Bouma, 1987) or volume (Mutti et al., 1984). For the present purpose, megabeds are defined by a thickness that exceeds 10 m.

Turbidites are commonly interpreted as lowstand deposits (e.g., Vail et al., 1977), but Mullins (1983) attributed deposition of calcareous turbidites to highstand periods. However, the deposition of megabeds has not been directly linked to sea-level lowstands or highstands. The preserved basin margin near Split in Dalmatia allows the study of the relationship between conglomeratic fan deltas and megabeds (Fig. 1).

DALMATIAN FLYSCH

A narrow belt of Cenozoic clastic rocks exposed along the eastern Adriatic coast is traditionally referred to as flysch of the Adriatic belt (Marinčić, 1981). It is made up of a stratigraphic span from Bartonian (Benić, 1983) to the late Miocene (de Capoa et al., 1991) in central Dalmatia (Fig. 2). In the studied area, flysch is of carbonate composition and is overlain by, as well as locally interfingers with, conglomeratic Promina Beds. The Dalmatian flysch includes the lower and upper flysch zone, divided by the K-S olistostrome (Marjanac, 1987a). These zones correspond to the synrift (lower flysch zone) and postrift (upper flysch zone) successions of the basin fill (Marjanac, 1991).

Megabeds occur in the lower flysch zone and the lowermost part of the upper flysch zone (Fig. 2) and form basinwide key beds (Fig. 1). In the central part of the basin they occur interbedded with (1) thin "classical" turbidites and (2) heterolithic units made of marlstones with thin, coarse-grained calcarenite interbeds, whereas at the northeastern basin margin they are interbedded with conglomeratic fan deltas (Figs. 2 and 3). Fan deltas occur in the upper part of the lower flysch zone and throughout the upper flysch zone. Several types of megabeds were recognized (Fig. 4) and will be reviewed only briefly below.

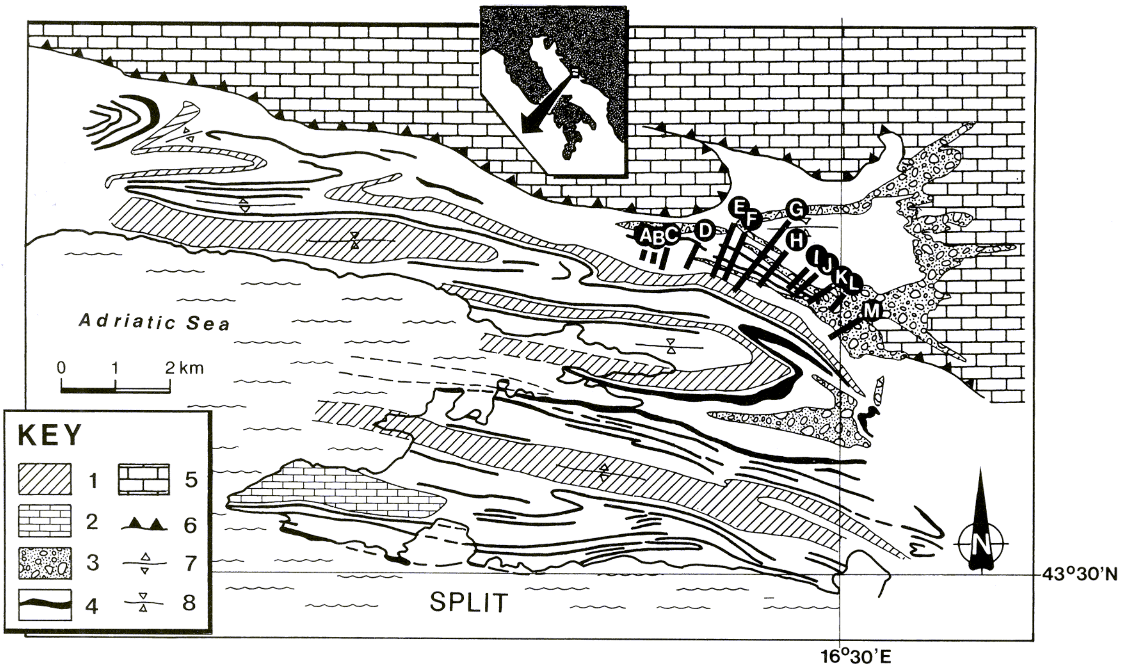

Figure 1. Map of Split environs modified after Kerner (1914) and locations of logs illustrated in Figure 4. Location of study area shown in inset map. Key: 1—K-S olistostrome, 2—Eocene Foraminiferal Limestones, 3—conglomerates, 4—megabeds, 5—Mesozoic carbonates, 6—thrust front, 7—axis of anticline, 8—axis of syncline.

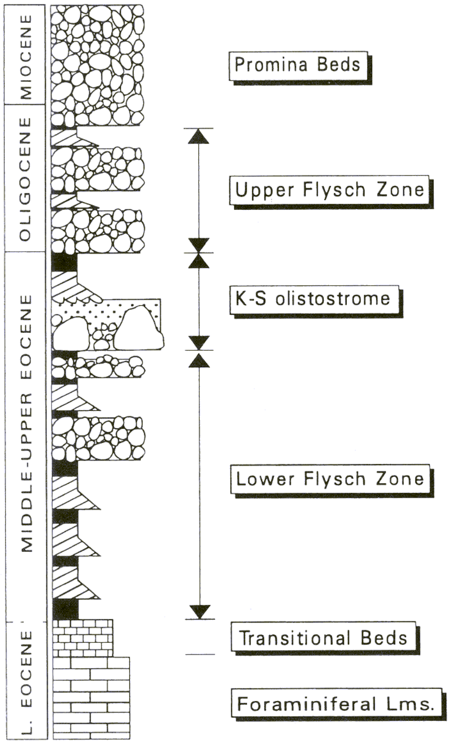

Figure 2. Schematic stratigraphic column of studied area. Stratigraphic boundaries are located only approximately. Megabeds alternate with thinner turbidites and marly heterolithic units in lower flysch zone (not shown owing to their small thickness) and with conglomerates in upper flysch zone.

TYPES OF MEGABEDS

Megaturbidites

Megaturbidites (Fig. 4A) are most frequent in the lower flysch zone, whereas only one such bed occurs in the lowermost part of the upper flysch zone (bed 188). They are as much as 27 m thick, are characterized by T_{a-e} and T_{c-e} Bouma sequences, and are composed predominantly of bioclasts and subordinately of lithoclasts.

"Reflected" Turbidites

"Reflected" turbidites (Fig. 4B) occur exclusively in the lower flysch zone and exceed 49 m in thickness. These beds were deposited from reflected and diverted flows, as indicated by alternating Bouma intervals and reversals of current structures at different levels within a bed. A more detailed account of this type of bed was given by Marjanac (1990). Although these beds seem to be rare in the ancient record, they have been reported from other areas (e.g., Hiscott and Pickering, 1984; Haughton, 1994).

Composite Turbidites

Composite turbidites (Fig. 4C) occur only in the lower part of the upper flysch zone (bed 2-7 in Fig. 3). They are composed of several lithologically uniform and normally graded, thinning-upward units and reach 37 m in thickness. These are a rare type of bed, and only a few have been reported in the literature (Cieskowski et al., 1994; Haughton, 1994; Marjanac, 1987b; van Straaten, 1971).

Complex Beds

Complex beds (Fig. 4D) are debrite-plus-turbidite (bipartite) beds. The thickest bed of this kind is the 170-m-thick K-S olistostrome (Marjanac, 1987a) with olistolites of Eocene Foraminiferal Limestones as large as 500 000 m^3 in volume, some of which are even karstified with small caves and pits. The turbidite part of some complex beds shows the internal complexity of the "reflected" turbidites.

The above megabeds are interpreted as products of high-density turbidity currents, surging turbidity currents, and cohesive debris flows. Deposition of megabeds is usually attributed to ponded turbidity currents of a high initial volume (Hiscott et al., 1986; Ricci Lucchi and Valmori, 1980). Triggering of these flows is usually attributed to seismic shocks (Kastens, 1984; Mutti et al., 1984; Séguret et al., 1984), but other causes have also been discussed (e.g., Mullins et al., 1986; Prior et al., 1982).

PROVENANCE OF DEBRIS

The studied megabeds are exclusively made of carbonate debris, both lithic and skeletal. The coarsest debris is represented by (1) extraclasts derived from the exposed hinterland (Eocene "Transitional Beds" and Foraminiferal Limestones; Aptian-Albian limestones), (2) "intrabasinal" clasts originating from erosion of semicontemporaneous calcarenites and marlstones, and (3) bioclasts (nummulitids and discocyclinids). All of these occur in the lower part of olistostromes, but Foraminiferal Limestones make up by far the largest clasts, some of which contain small caves. Since olistolites were encased in marly matrix before weathering exposed them, this makes an unlikely condition for in-situ speleogenesis. So, it seems more reasonable to assume that their karstification occurred *prior* to redeposition.

Very abundant foraminiferal bioclasts are identical to the fauna of lithoclasts. Some nummulitids in matrix of the K-S olistostrome are only partly washed out from Foraminiferal Limestones, indicating their origin by weathering of these limestones.

Composition of lithoclasts indicates an External Dinaric provenance, which suggests uplifting, subaerial exposure, and intensive weathering already during the Eocene and (possibly) Oligocene.

SEA-LEVEL CHANGE AND DEPOSITION OF TURBIDITES

Turbidites are commonly attributed to periods of relative sea-level lowstands, when terrigenous debris reached directly to the shelf break (e.g., Vail et al., 1977; Shanmugam and Moiola, 1982, 1984). However, carbonate turbidites have sometimes been attributed to sea-level highstands, when productivity of platform carbonate mud was at a maximum (Dolan, 1989; Masetti et al., 1991; Mullins, 1983). Piper and Savoye (1993) found the Holocene (highstand) turbidites to be sandier, with more abundant skeletal debris, than the Pleistocene muddy turbidites, but also thinner than their Pleistocene counterparts (van Straaten, 1971). Dolan (1989) reinterpreted turbidites of the Deep Sea Drilling Project (DSDP) 48 site (previously interpreted as of a lowstand origin, Thiede, 1981) and attributed them to the initial stage of sea-level lowering, just prior to exposure of the platform. Reymer et al. (1990) found the Bahamian platform–derived lowstand turbidites to be rich in platform-interior–derived debris and the highstand turbidites to be rich in platform-margin debris. Tunis and Venturini (1992) attributed massive siliciclastic turbidites to lowstand phases and "allodapic" limestones and/or fine siliciclastic turbidites to highstands.

Generation of megabeds has seldom been attributed to a specific sea-level stand, probably because they are usually related to tectonic (seismic) events that "totally mask the eustatic effect" (Tunis and Venturini, 1992, p. 144). Shanmugam and Moiola (1984) attributed initiation of major turbidity currents (such as the Roncal Unit of the Spanish Pyrenees) to the initial stage of sea-level lowering. Similarly, Yose and Heller (1989, p. 437) attributed major collapses of outer ramps and upper slopes to "impinging of storm wave base . . . when the rate of sea-level fall has increased to a maximum (lowstand . . .)." However, Ricci Lucchi (1990) reported great thicknesses of highstand turbidites, some of which were also reflected(!).

DISCUSSION

The incipient construction of coarse-grained fan-delta bodies is related to sea-level falls and lowstands, whereas change from progradation to aggradation (Fig. 3) may be explained in terms of variation in generation of the new accommodation space as a function of sea-level fall and subsidence. Some of the studied fan-delta conglomerates interfinger with turbidites (Fig. 3), suggesting that these have also been deposited during relative sea-level lowstand. However, these turbidites are one order of magnitude thinner than megabeds, suggest-

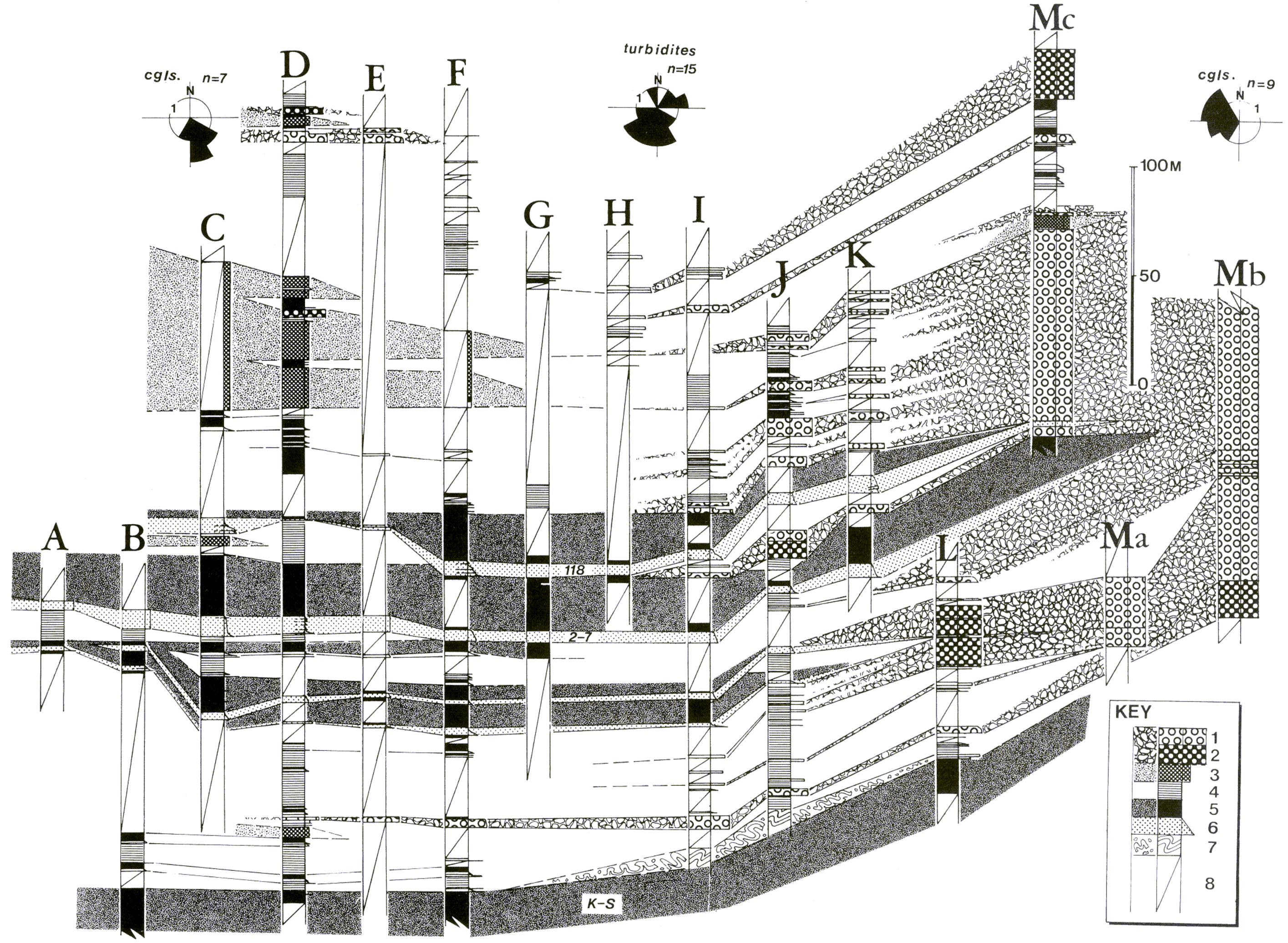

Figure 3. Schematic west-east section of western Mosor Mountain, ~4 km long; locations of logs shown in Figure 1. Note position of megabeds (2-7, in particular) between major progradational fan-delta conglomeratic deposits and interfingering of megabeds with thinner turbidites. Current roses for conglomerates of western and eastern provenance, as well as turbidites, illustrate different paleotransport modes, even some upslope turbidity flows. Key: 1—conglomerate, 2—debrite, 3—sandstone, 4—heterolithic units (marlstones with thin calcarenite interbeds), 5—turbidite marlstones, 6—turbidite arenites, 7—slump, 8—covered section.

ing a higher frequency of flows and comparatively smaller volume.

Generation of megabeds is usually attributed to catastrophic collapses of basin margins, and initiation of large-scale gravity flows of sediment fed by linear sources seems less likely during sea-level lowstands for several reasons: (1) Wave action does not commonly trigger catastrophic collapses and flows, because the frequency of large storms is much higher than that of megabeds. Recurring periodicity of the early Holocene turbidites in the Mediterranean Sea is about 1500 yr (Kastens, 1984), and that of the Paleogene megabeds of the southern Pyrenees is about 500 000 yr (Séguret et al., 1984). (2) Sea-level fall exposes the shelf, which is then desiccated, causing initial compaction, which increases the mechanical stability of slopes. (3) Early (vadose) cementation would further increase stability of the slope during the exposure.

Being infrequent events, the deposition of megabeds must have occurred after significantly long periods of primary accumulation of debris to account for their large volumes (provided there was no significant variation in debris supply). A large quantity of lithic and weathered skeletal debris implies long-term weathering of the emergent land, presumably during periods of relative sea-level lowstands and (possibly) transgressions.

During a rise in sea level, the ground-water table rises, tides amplify, and wave action increases. Pore-water pressure will increase and may provide favorable conditions for slope instability and eventual collapse of the slope. In this way, linear sources along basin margins may occasionally generate megaflows. Favorable conditions for the catastrophic slope failures seem to be met when the rate of sea-level rise is highest—a time when generation of maximum flooding surfaces is expected.

Megabeds that are interbedded with fan deltas (Fig. 3) were deposited when the deposition of coarse-grained debris had shifted landward, that is, during rising sea level. This conclusion is supported by the following facts: (1) megabeds were deposited during periods of aborted supply of coarse-grained debris, as their coarsest debris is much finer grained than the average clast size in underlying conglomerates (except for the K-S olistostrome), (2) megabeds are overlying some fan deltas, (3) since most of

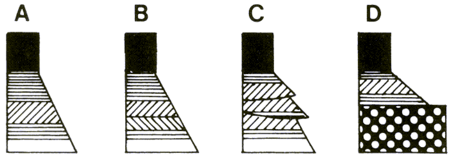

Figure 4. Types of megabeds recognized: A—megaturbidites, B—"reflected" turbidites, C—composite beds, D—complex beds.

the bioclasts are resedimented, the flows must have occurred after a long period of weathering to account for their large quantity, and (4) compositional difference indicates that fan deltas did not supply debris for megabeds.

The K-S olistostrome was formed by the catastrophic collapse of the shelf margin and adjacent exposed land, as indicated by the karstified olistolites. It is very probable that this event occurred during sea-level rise to account for shaping of olistolites during lowstand and production of the large quantity of skeletal debris by weathering of older strata (mainly Foraminiferal Limestones). The resedimented skeletal debris makes biostratigraphic dating extremely difficult. Hence, it was not possible to estimate duration of the lowstand subaerial exposure, but it must have lasted long enough for karstification to generate small caves.

CONCLUSIONS

Whereas relatively thin, "classical" turbidites (interfingering with fan-delta conglomerates) have obviously been deposited during relative sea-level lowstands, the megabeds that are interbedded with fan deltas document deposition during sea-level rise. Initiation of megaflows sometime after a lowstand (presumably during accelerated sea-level rise) is indicated by the karstified olistolites and the abundant skeletal debris, which originated by subaerial weathering of the Foraminiferal Limestones. Megabeds found in a basinal setting of the lower flysch zone may also correspond to events of the sea-level rise, whereas other terrigenous sediments (turbidites, thin-bedded turbidites, and sandy marls) were probably deposited during relative sea-level lowstands. Discussed megabeds originated by shelf-margin collapses, and their great volumes may be attributed to long periods of primary accumulation of debris and lower frequency of these catastrophic events. The response of linear sources (megabed-generating shelf margins) to sea-level changes is thus different from that of point sources ("classical" turbidite-generating settings).

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