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Condensed Sections: Terminology, Types, and Accumulation Conditions

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Abstract—Condensed sections (CS) are abnormally thin but nominally complete sections embracing large stratigraphic intervals. They are subdivided with regard to sedimentation rate into CS (at sedimentation rate of 0.5–1 cm/ka) and ultracondensed sections (with a sedimentation rate <0.5 cm/ka), which can be both deep-water and shallow-water types. The mechanisms of CS accumulation include mechanical, chemical, biotic condensation, and nondeposition. Their combinations allow one to suggest some models of CS accumulation: (1) fast and (2) slow sea level rise and/or its low-amplitude fluctuations; (3) sea level drop; (4) global warming and (5) global cooling.

Key words: hiatus, condensed section, sedimentation model.

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INTRODUCTION

Condensation is understood in stratigraphy and sedimentology as a sharp decline in sedimentation rate and accumulation of a thin succession embracing a broad stratigraphic interval. Condensed sections (CSs) are very common in sedimentary basins of various types, and their investigation is important for various reasons. They mark significant events in the geological evolution of sedimentary basins related to tectonic, paleogeographic, and sedimentation rearrangements. CSs are reliable regional correlation bench marks applicable in stratigraphy and seismic stratigraphy. Phosphorite, glauconite, and manganese deposits, as well as elevated concentrations of PGM and other elements, are confined to condensed sections [Loutit et al., 1988].

Notwithstanding the fact that the terms "condensation," "condensed units," and "condensed sections" are common in the geological literature, these terms often imply very different sedimentation environments and a broad variety of successions. The present author undertakes an attempt at ordering the usage of these terms and at briefly reviewing the principal CS types and possible mechanisms of their formation.

DEFINITION

Most publications define condensed sections as abnormally thin successions corresponding to long time intervals. A classical publication on CSs [Loutit et al., 1988] defined them as "...thin marine stratigraphic units consisting of pelagic and hemipelagic sediments of which extremely low sedimentation rates are characteristic." In this case, CS formation is usually associated with rapid sea level rise [Loutit et al., 1988; Vail et al., 1984].

In my opinion, this definition contains some shortcomings. First, it does not take into account the variety of CS formation environments, viz., CSs are as common in shallow-water environments [Baraboshkin et al., 2002; Gomez and Fernandez-Lopez, 1994; Hillgaertner, 1998]. Second, sections that enclose fossil complexes, which occurred due to redeposition because of erosion and not due to low sedimentation rate, were also included into condensed sections.

In view of the above stated, condensed sections are **abnormally thin but nominally complete sections**, i.e., those enclosing all (or almost all) subdivisions of the global stratigraphic scale of the corresponding rank, **embracing large stratigraphic intervals** (several biostratigraphic zones, a substage, or a stage) **that originated due to a sharp decline in the sedimentation rate, whose accumulation was interrupted by intervals of nondeposition, erosion, and other synsedimentary or early diagenetic events** [Baraboshkin et al., 2002]. The types of synsedimentary and early diagenetic hiatuses are shown in the table. This definition of a CS is close in its understanding to those proposed by E. Rod [Rod, 1946] and H. Jenkyns [Jenkyns, 1971].

It is evident that the term "condensation horizon", which also frequently occurs in the literature and is common in geological terminology, is not a synonym of a CS as a layer enclosing a redeposited ("mixed") fossil complex [Baraboshkin et al., 2002; Jenkyns, A group of synsedimentary and early diagenetic hiatuses, modified after [Baraboshkin et al., 2002]

Hiatuses		Principal process		Sedimentation type		
				carbonate		terrigenous
Type	Kind			shallow-water	deep-sea	shallow-water
Nondeposition	Nondeposition	Nondeposition (s.s.)	Long-term	Lack of sediment		
	Diastems		Short-term			
Erosional	Erosional	Sediment erosion	Complete	Erosional surfaces		
			Noncomplete	Conglomerates (residual sediments), intraclasts		
Underwater	Eluvial–diagenetic (hard bottom [HB], concretions, knotted chalk)	A set of synsedimentary and early diagenetic factors		Beachrock, HB, knotted chalk	Iron–manganese crusts and nodules	Phosphorite crusts
				Phosphorite and carbonate slabs		Phosphorite, siderite, and other slabs
	Soft bottom	Bioeluvium		Soft bottom	Soft bottom (rarely)	Soft bottom
	Halmyrolytites	Chemoeluvium		Glauconite	Bentonite	Glauconite
		CCD displacement		Clay interlayers	Red oceanic clay	–
		Cooling/biological productivity decline				
Subaerial	Expositions	Various synsedimentary and early diagenetic				
	Soils (paleosoils)					

1971]. I have to mention here that there are two terms describing different types of redeposition: "redeposition" proper (resedimentation or desedimentation), i.e., redeposition of more ancient rocks and "reelaboration," i.e., almost synsedimentary reworking of sediments. The first phenomenon is attributable to condensation horizons, while the second, to the condensation process proper.

The term condensation horizon is close in its meaning to the term "condensation surface" [Fluegel, 2004; Hillgaertner, 1998].

In some publications, for example, in [Fluegel, 2004], the term condensation is a synonym of "concentration." Condensation and concentration are different phenomena, although some researchers do not separate them. Sedimentary concentration is a process of accumulating some specific objects (phosphorite, siderite, or cherty nodules, magnetite grains, fossil organic remains, etc.) in a part of a sedimentary basin; and this process is not related to stratigraphic stratification of sediments. The principal difference between condensation and concentration is precisely that despite the small thickness no mixing of sediments of different ages takes place (or almost does not take place), i.e., the primary stratification of sediments is preserved. They differ in this respect from, for instance, cherty conglomerates, where ancient and young cherts are mixed as pebbles and therefore are condensed (this can be observed in coastal precipices on Rügen Island).

Rate of CS accumulation. I have to stipulate from the very beginning that only average values of sedimentation rate are discussed since, as G. Einsele [Einsele, 2000] rightfully noted, absolute value of sedimentation rate cannot be calculated in most cases.

What sedimentation rate can be regarded as low? No precise answer to this question exists, and the scatter of suggested values is rather wide. Some researchers regard a section as condensed if the sedimentation rate reaches 16 cm/ka. However, most authors estimate the sedimentation typical of a CS as considerably lower. P. Vail and his coauthors [Vail et al., 1984], particularly, estimated the rate precisely, viz., below one cm/ka.

It is evident that sedimentary material sources are different under different paleogeographic conditions. A low sedimentation rate in shallow-water basins is caused by the decrease or termination of clastic and/or biogenic (at low latitudes) material supply. The sedimentation rate in deep-water parts of a sedimentary basin is determined by the supply rate of background pelitic and mainly planktonic sediments.

If we follow the definition of a CS suggested in several publications [Loutit et al., 1988; Vail et al., 1984], then we have to accept the rate of accumulating pelagic and hemipelagic sediments as a guide to accumulation rate. The lowest rate is typical of pelagic areas of oceans (0.1–6 cm/ka) [Kukal, 1987; Lisitsyn, 1988; Romanovskii, 1998; Einsele, 2000]; in this case, the lowest values of sedimentation rate were recorded both in red oceanic clay (usually below 5 mm/ka) and in deep-sea radiolarite [Kukal, 1987; Reding et al., 1990]. Similar rate values were recorded in limestone of the *Ammonitico Rosso* facies [Baraboshkin et al., 2002; Reding et al., 1990; Einsele, 2000] in some CSs of flooded carbonate platforms in the Mediterranean [Reding et al.,1990, in CSs of epicontinental terrigenous basins [Baraboshkin et al., 2002], and in Silurian graptolite shale [Reding et al., 1990]. According to G. Einsele [Einsele, 2000], the same sedimentation rate is characteristic of biogenic sedimentation areas (bituminous shale, pelagic carbonates, isolated deep-sea carbonate platforms, radiolarite, and siliceous ooze) and on bypassing shelves.

The values of 0.5 and 1 cm/ka are some threshold values characterizing sedimentation rates typical of CSs [Kukal, 1987; Lisitsyn, 1988; Romanovskii, 1998; Einsele, 2000; Loutit et al., 1988; Vail et al., 1984]. The present author suggests distinguishing a CS proper (with an average sedimentation rate of 0.5–1 cm/ka) and ultracondensed sections (UCS with an average sedimentation rate < 0.5 cm/ka).[1] This conditionality refers, naturally, to average rate values, since they are sums of episodes of complete nondeposition, erosion, and episodes of relatively fast sedimentation often exceeding 1 cm/ka.

CS types. T. Loutit and his coauthors [Loutit et al., 1988] stated that a CS is always associated with the maximum depth of a basin. It was found later that it is not always so [Baraboshkin et al., 2002; Gomez and Fernandez-Lopez, 1994; Hillgaertner, 1998]. Based on published data and from this author's personal observations, all CSs can be most generally subdivided into the following types:

1. Shallow-water[2] (neritic shoals and the inshore parts of a basin with deficient sediment supply and those that formed due to expansion of pelagic sedimentation).
2. Hemipelagic and pelagic shallow-water types (epicontinental basins; shallow-water submarine uplifts without a supply of terrigenous material, submerged isolated platforms; and shelves subjected to the action of oceanic currents).
3. Hemipelagic (bypassing basin slopes).
4. Pelagic deep-water (abyssal, deep-sea uplifts).

Shallow-water CSs include the following types of sections:

(1) Shallow-water sediments of the Early–Middle Ordovician ramp in the Russian Plate (Fig. 5). The presence of glauconite, layers of micronodules, and numerous hard bottom surfaces are characteristic of these CSs. This CS type originated under conditions of limited (deficient) supply of sediments (clastic and carbonate) due to the action of flood-ebb currents [Zaitsev and Baraboshkin, 2006].

(2) Mixed carbonate–terrigenous Lower Valanginian and basal Upper Valanginian sediments in the South West Crimea [Baraboshkin et al., 2002]. The rocks consist of bioclastic limestone, sandstone, clay, and "pudding" conglomerate enclosing phosphorite and iron oolites. They accumulated under conditions of the beginning of an Early Cretaceous regional transgression in nearshore shoals during active action of sea heaving, currents, and biological activity.

G. Hillgaertner [Hillgaertner, 1998] noted the possibility of carbonate nearshore CS formation because of the fast cementing of numerous hard bottom layers due to the action of flood–ebb currents, even without sediment deficiency.

Hemipelagic and pelagic shallow-water CSs include a great spectrum of sections:

(1) Shallow-water pelagic (globigerine) limestones on Malta Island, which are planktonic organically disturbed biological micrites enclosing glauconite, phosphorite interlayers, echinoderms, mollusks, and ostracodes. They accumulated on a submarine uplift that was isolated from the terrigenous material supply [Reding et al., 1990].

(2) Paleozoic and Mesozoic red-colored and variegated biomicritic cephalopod limestones ("Orthoceratitico Rosso," "Goniatitico Rosso," and "Ammonitico Rosso" facies) in many sections in Eurasia and Africa (particularly, in the Tethys Paleobasin) that often overlie shallow-water carbonates on within-basin uplifts (Fig. 2). The condensation occurred due to almost simultaneous pelagic sedimentation and the action of near-bottom currents, as well as submarine dissolution, cementing, organic erosion, and encrustation, which were accompanied by the sedimentation of iron and manganese hydroxides [Baraboshkin et al., 2002; Lisitsyn, 1974; Reding et al., 1990; Jenkyns, 1971]. The depth at which the cephalopod limestone accumulated varied between 1000 m and 0 m, though most researchers agree that it hardly exceeded 200–300 m [Reding et al., 1990]. I think that the bathymetric range of cephalopod limestone accumulation was rather wider and therefore only some of such CSs belong to this group and some of them belong in the group of pelagic CSs (*s.s.*).

(3) Some Mesozoic bituminous shales ("black shales") in epicontinental basins that accumulated during transgressions (consequently, during the expansion of pelagic basins) and consist of background sediments (clay) saturated with organic matter. They originated under conditions of high planktonic biological productivity during periodic or permanent anoxia [Reding et al., 1990]. Their accumulation rates vary strongly (0.5–10 cm/ka, [Einsele, 2000]), which only partly corresponds to the rates of CS accumulation. Therefore, only some bituminous shales are CSs.

(4) "Glauconite marl" lies at the base of writing chalk successions in southwestern England, in the Caucasus, in the Transcaspian Region, in the Russian Plate, and in other areas and marks the transition from

[1] The term "ultracondensation" was applied earlier to a glauconite–phosphatic CS in Aptian–Cenomanian sediments in Escragnolles (Southeastern France) [Cotillon, 1983].

[2] Shallow-water in this case means shallow-marine environments with a depth that does not exceed the average depth of a contemporary shelf, i.e., 200 m.

shallow-water marine clastic sediments to pelagic environments. According to [Reding et al., 1990], this is an example of ancient pelagic facies. The rock is phosphate-bearing organically disturbed chalk enclosing abundant and variable phosphatized and nonphospahatized fossils (sponges, mollusks, brachiopods, sea urchins, solitary corals, etc.) and phosphorite nodules. Similar sediments occur on the Blake Plateau near the Californian Coast and on the Agulhas bank at the South African Coast [Reding et al., 1990; Shopf, 1982]. Redeposition of phosphorites and organic remains, plankton abundance, and authigenic mineralization were caused by cold bottom currents that rose from the depths nearly to the surface.

(5) Numerous epicontinental sections of glauconite–phosphorite Phanerozoic sandstones. These CSs differ from the previous type only by their considerably lower carbonate abundance. The most typical include Aptian and Albian (Gault) sections in Western Europe, Volgian Stage, Berriasian–Valanginian [Baraboshkin et al., 2002] (Fig. 4), lower and middle Albian in the central part of the Russian Plate [Baraboshkin et al., 2002], Barabinsky Member (upper Oxfordian–Kimmeridgian) in West Siberia (Fig. 1), and many others. These CSs originated under conditions of a rapid sea level rise and under the action of waves, currents, and organic activity. The origin of authigenic minerals is due to an increase in the biological productivity in the basin caused by expansion of the pelagic basin and by the action of shallow-water upwellings or currents that brought cold water masses from the Boreal Basin.

(6) Sections of Aptian–Albian glauconite–phosphorite sandstones on flooded shallow-water carbonate shelves in Tethys Basin (Fig. 3).These CSs are very similar to the previous type, but they accumulated within the passive margins of the Tethys Ocean [Reding et al., 1990]. The principal factors that caused their formation included increases and oscillations of the sea level and storm/wave activity that caused washout of fine sediments; these included also the action of cold currents. Glauconite–phosphate sedimentation was possibly related to the large supply of nutrients into the basin (eutrophication), which is partly supported by its correlation with global anoxic events [Shopf, 1982; Einsele, 2000].

The carbonate sections on guyots are similar in some respects to this type [Reding et al., 1990].

Hemipelagic CSs originated on the bypass slopes of deep-water basins [Reding et al., 1990; Einsele, 2000] in those places where the sediments were comparatively rapidly transported toward the continental rise. The Bahamas Bank (Caribbean Basin) is a present-day example of such a CS.

Pelagic deep-sea sediments include the following types:

(1) Cephalopod limestones (see above) that accumulated on deep-sea uplifts (deeper than 200 m).

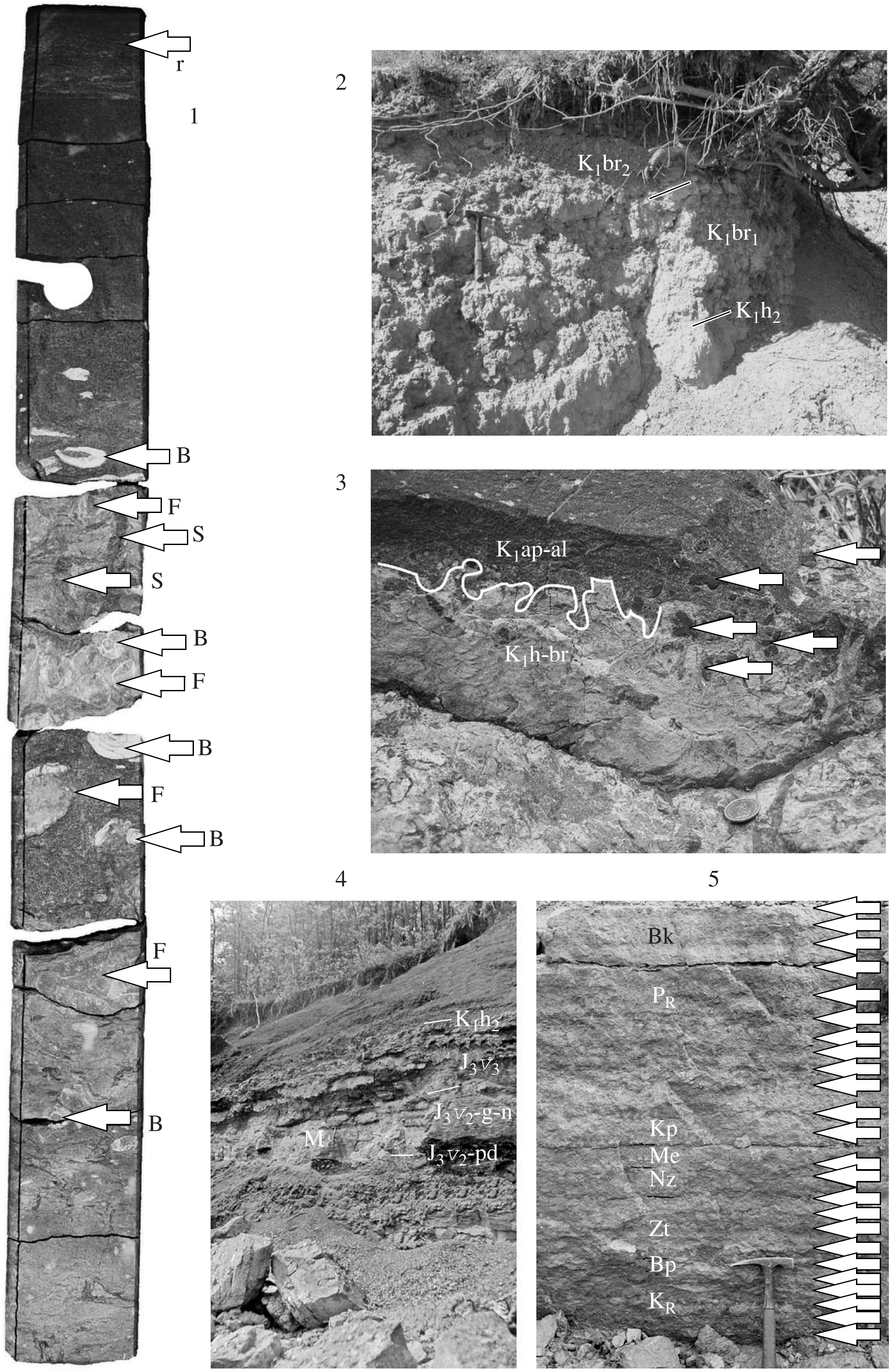
(2) Red oceanic clay, which are (with a few exceptions) UCSs. Sediments of this type cover vast areas (up to a quarter of the present-day ocean floor), and they accumulated below the carbonate compensation depth (CCD) within the abyssal zone. The specific composition of the clays related to the great distance from source areas and extremely low sedimentation rate allowed one to call them polygenic and permitted the separation of several types among them (transitional, eupelagic, and zeolitic) [Lisitsyn, 1974]. The red Cretaceous clay on Timor Island presents a classical fossil section of such sediments [Reding et al., 1990].

(3) Siliceous (diatomaceous and radiolarian) pelagic sediments that accumulated below the CCD and are present in ophiolite successions. These CSs more rarely consist of planktonic carbonates, since the rate of their accumulation is rather high (particularly, in zones of oceanic divergence). Siliceous and siliceous–phosphate sedimentation takes place currently in areas of upwelling along the western margins of continents where the sedimentation rate is higher [Lisitsyn, 1974, 1988]. Along with this, the number of hiatuses is also greater along the western margins of continents where they are also caused by currents

Some examples of condensed sections: Panel 1. Shallow-water terrigenous glauconite–phosphorite CS. West Siberia, Tomsk oblast, West Moiseevsky-30 Well, Barabinsky Member (Oxfordian–Kimmeridgian) and its boundary with clay of the Georgievsky Formation (Kimmeridgian, Lower Volgian Substage, G). CS consists of organically disturbed glauconite–quartzose sandstone enclosing phosphorite (F), belemnite rostra (B), and *Scolithos* organic disturbances (S). The core diameter is 10 cm. Panel 2. Shallow-water hemipelagic CS. Cephalopod limestone ("Ammonitico Rosso," upper Hauterivian–Barremian) on the southern slope of Mount Belaya, right slope of the Kacha River valley, South West Crimea. Indexes designate: K_1h_2 is upper Hauterivian; K_1br_1 is lower Barremian; and K_1br_2 is upper Barremian (*Heinzia provincialis* Zone). Panel 3. Shallow-water hemipelagic CS. Mount Pilat area, Helvetian Alps, Switzerland. Boundary between Hauterivian–Barremian nearshore grainstone of flooded shallow-water carbonate platform (K_1h–br) and the CS base of Aptian–Albian (K_1ap–al) glauconite–quartzose phosphorite-bearing sandstone. Arrows indicate the largest phosphorite nodules. Coin diameter is 2 cm. Panel 4. Shallow-water terrigenous glauconite–phosphorite CS. Right bank of the Volga River valley near the Undory Settlement (area of the Volgian Stage stratotype). Noncondensed portions of the succession: J_3v_2–pd are clay and bituminous shale of mid-Volgian Substage with numerous eluvial hiatuses (*Dorsoplanites panderi* Zone). K_1h_2 is upper Hauterivian clay (*Speetoniceras versicolor* Zone). The CS consists of organically disturbed glauconite–quartzose sandstone with phosporite condensation horizons J_3v_2–g–n–J_3v_3 (= K_1brs) (mid-Volgian Substage, *Virgatites gerassimovi–Epivirgatites nikitini* zones are upper-Volgian Substage (= lower Berriasian). M is hammer. Panel 5. Shallow-water carbonate CS. Leningrad oblast, a quarry at the Babino Settlement. A fragment of the succession of the Volkhov Unit (Lower Ordovician). Arrows indicate positions of hard bottom surfaces that originated due to tide-and-ebb currents. Hammer length is 35 cm.

[Kennett, 1987]. One cannot rule out the possibility that CSs originate even in upwelling areas.

In addition to a sedimentary origin, thin sections can accumulate during deep diagenesis and after its completion. Such sections are "pseudocondensed" and are not related to real CSs. Examples of the first are intervals of carbonate successions with numerous stylolite surfaces and with great amplitude of stylolites



(for example, Turonian sediments in the central part of North Caucasus). Examples of the second are evaporite successions where thick salt layers "flow out" of intervals between limestone, dolomite, and other rock beds [Raevskii et al., 1973], and these beds mechanically close up and become condensed.

Mechanisms of CS formation. Some authors discriminate several types of condensation for CS formation: stratigraphic, sedimentational, and taphonomic [Fuegel, 2004; Gomez and Fernandez-Lopez, 1994]. Stratigraphic condensation is understood, after [Heim, 1934], as a decline in sediment supply, up to its complete termination. Sedimentational condensation implies a decline in the sediment accumulation rate. Taphonomic condensation is understood as the mixing of organic remains of different ages.

All types of condensation present in successions are naturally stratigraphic (although the causes of their formation can be very different), and therefore separation of stratigraphic condensation as an independent type is senseless. The two first types are associated with CS formation and are known as nondeposition and hiatus formation (synsedimentaton erosion of different kinds, biological erosion, dissolution, early diagenesis, etc. [Baraboshkin et al., 2002], (Table)), in this case, the second, i.e., hiatus formation often follows from the first (if there was no sediment, it did not accumulate). The difference between these terms consists, in the opinion of some authors (particularly, [Gomez and Fernandez-Lopez, 1992]), in that there are cases when sediment does not accumulate, even with a large material supply (for example, carbonate dissolution below the CCD), although episodes of high accumulation rate more often alternate with nondeposition. In accordance with the definition, condensation always takes place during sedimentation and early diagenesis and therefore it is usually synsedimentary. I think that it is more justifiable to determine the stages during which condensation takes place (synsedimentary and early diagenetic) and a concrete condensation mechanism.

I do not regard taphonomic condensation as being very common at CSs and particularly condensation horizon formation as condensation proper, since, during this type of condensation, first, the most important indication of condensation, i.e., preservation of the original stratification is not fulfilled and, second, usage of the term "taphonomy" is limited to the realm of organic objects.

The principal mechanisms of CS formation include the following (modified and supplemented after [Fluegel, 2004]):

1. *Mechanical condensation includes*: erosion, washing out of fine material, concentration and redistribution of coarse material under conditions of active hydrodynamics; mechanical precipitation of background or nannoplankton sediments under conditions of weak hydrodynamics or in still water; and compaction of sediment. Many researchers view currents acting at all depths [Lisitsyn, 1988; Hillgaertner, 1998; Jenkyns, 1971] and wave action [Hillgaertner, 1998] as the leading process of condensation. The importance of the gravity sloughing/fall of sediments from slopes of inner uplifts is insignificant.

2. *Chemical and biochemical condensation includes*: dissolution and mineral replacement; growth of authigenic minerals typical of CSs (glauconite, phosphorite, iron–manganese crusts and nodules, etc.). This condensation is particularly important in oceans where carbonates dissolve below the CCD.

3. *Biotic condensation includes*: eating, organic disturbance, biotic erosion, encrustation, biotic redistribution and compaction of sediment. This condensation mechanism is most common in shallow-water areas.

4. *Nondeposition* when sediment does not reach the seafloor. Nondeposition may have different causes: transport of solid particles in suspension, dissolution of sedimentary particles, consumption of organic detritus in the water column, etc.; nondeposition takes place during all processes mentioned above and may be the result of a single or several processes. Nondeposition opens the way to all other mechanisms, and therefore it is regarded as an independent mechanism.

Different combinations of the above-named factors bring about CS formation.

Models of CS formation. Concrete occurrence and combinations of condensation mechanisms in different paleogeographic environments allow reviewing the following models of CS formation:

(1) rapid sea level rise;

(2) slow sea level rise and/or its low-amplitude oscillations;

(3) sea level abatement;

(4) global warming and displacement of convergence and divergence zones;

(5) global cooling and displacement of convergence and divergence zones.

Rapid sea level rise (relative; its causes are not discussed here; it is caused by a combination of the rate and direction of eustatic and tectonic motions, basin filling, etc.) brings about the transformation of shallow-water and nearshore sections into pelagic ones, flooding of lowlands, decline in the erosion rate, and a decrease of erosion area. This is, naturally, the principal reason for CS formation [Lisitsyn, 1988; Gomez and Ferandez-Lopez, 1994; Jenkyns, 1971; Loutit et al., 1988; Vail et al., 1984]. Shallow-water hemipelagic and pelagic CSs originate during rapid sea level rise under shelf or epicontinental basin conditions. Anoxic pelagic deep-sea and shallow-water epicontinental CSs can also occur [Einsele, 2000]. Even a low sea level rise in an epicontinental basin with a thin sedimentary cover and low topography is sufficient for flooding sources of sedimentary material supply and for CS formation [Baraboshkin et al., 2002].

In deep-sea pelagic environments, a sea level rise causes CS formation, and a sea level drop brings about the formation of a hiatus surface; the lithologic composition of CS can also change [Reding et al., 1990].

Deep-sea pelagic UCSs are different: a rise in sea level causes sediment and biophile element concentration at shelf depths, which brings about a reduction of the biogenic sedimentation rate in the pelagic region (hiatuses originate) [Lisitsyn, 1988].

Seafloor topography is an additional factor of CS formation in all the above-mentioned cases [Lisitsyn, 1988]; the CS area is closely related to the area of within-basin uplifts [Reding et al. 1990].

A rise in sea level is accompanied by an upwelling of cold oceanic water on the shelf. This process results, on one hand, in the formation of a shallow-water hemipelagic CS with glauconite and phosphorite and, on the other hand, in an additional decline in sedimentation rate due to sediment rewashing. The Aptian–Albian CS on flooded carbonate platforms in the Mediterranean may have formed in precisely this way.

A slow rise in sea level and/or its low-amplitude oscillations on carbonate shelves bring about the formation of shallow-water and hemipelagic shallow-water CSs. They originate due to preponderant transport of sediments into deeper-water regions of the basin by ebb currents and fast cementing of carbonates with the formation of hard bottom surfaces [Zaitsev and Baraboshkin, 2006; Hillgaertner, 1998].

A decline in sea level brings about the mass transport of sediments to the continental rise and increased sedimentation in the deep-sea pelagic region [Lisitsyn, 1988]. CSs in shallow-water areas do not form during such periods, since erosion processes intensify and hiatus surfaces form. Biological productivity grows in the pelagic areas, which brings about CS formation at great depths [Lisitsyn, 1988] and rapid sedimentation in UCS areas.

Global warming and displacement of convergence and divergence zones cause anoxic phenomena and CSs related to them. The main event related to these phenomena is a decrease in the CCD [Kennett, 1988; Lisitsyn, 1974], which gives rise to “diffuse silica accumulation” [Lisitsyn, 1988] and a significant increase of the area of CS and UCS formation. A rise in sea level and intensification of an upwelling system may be related to global climate warming, but are probably not always related to it.

The displacement of convergence and divergence zones correlates with changes in atmospheric conditions. This phenomenon does not necessarily cause changes in the area of CS formation but it can change the area’s location. The combined effect of the CCD decrease and displacement of convergence and divergence zones can increase CS and UCS areas.

Global cooling and displacement of convergence and divergence zones. Cooling gives rise to a decrease in the CCD and to a decline in the rate and area of carbonate production (both neritic and pelagic). Simultaneously, it increases the area and rate of biogenic silica production [Lisitsyn, 1988]. Therefore, cooling causes CS formation in shallow-water areas and a decrease in the area of a deep-sea pelagic UCS. Cooling intensifies erosion, upwelling processes, and the oceanic current system, which brings about active accumulation of contourites in abyssal areas and terminates deep-sea anoxia [Lisitsyn, 1988; Shopf, 1982]. Sea level decline, erosion in shallow-water zones, and displacement of accumulation area into the bathyal zone often accompany climatic cooling.

A displacement of convergence and divergence zones during climate cooling does not result in increasing the area of deep-sea pelagic CSs and UCSs.

The tectonic setting of the sedimentation area is the most important factor of shallow-water CS formation: the average sedimentation rate in epicontinental basins is lower by 1–2 orders of magnitude than in shallow-water and marginal seas [Shopf, 1982].

The location of a continent at high latitudes or, in other words, the width of arid climate zones, is an additional factor for the origin of any CS [Shopf, 1982]. According to Lisitsyn [1988], the equatorial zones supply the ocean with three times more sedimentary material than all the other zones of the Earth. Therefore, conditions for CS formation were most favorable during those periods when the continents were located at high latitudes (for example, during the Early Paleozoic).

CONCLUSIONS

It follows from the above data that CSs and the causes of their formation are much more variable than previously presumed [Loutit et al., 1988; Vail et al., 1984]. They were related to global fluctuations of the sea level, to its declines, and to climatic rearrangements. The sedimentary system reacts differently to these processes under different conditions, and therefore, CS and UCS formation was most active in shallow-water areas at some times and in deep-sea zones at other times. This should be taken into consideration during stratigraphic correlations. It is evident that evolution of the biota, first of all, plankton is an additional factor of CS formation. For this reason, CSs and UCSs that originated during different epochs of the Earth’s evolution differ greatly from each other.

As a rough approximation, the rate of clastic material supply (particularly, at high and middle latitudes) and the rate of neritic carbonate formation (at low latitudes) determine CS origin in shallow-water areas. CS formation in deep-sea areas is almost exclusively related to the rate of biogenic (planktonic) sedimentation, which, in turn, depends on biophile element supply and on the Earth’s climatic conditions.

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