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A MINERALOGICAL AND GEOCHEMICAL APPROACH TO ESTABLISHING A SEDIMENTARY MODEL IN A PASSIVE CONTINENTAL MARGIN (SUBBETIC ZONE, BETIC CORDILLERAS, SE SPAIN)

M. ORTEGA-HUERTAS, I. PALOMO, M. MORESI* AND M. ODDONE†

*Departamento de Mineralogía y Petrología, Instituto Andaluz de Geología Mediterránea, Universidad de Granada, Spain, * Dipartimento Geomineralogico, Università di Bari and † Dipartimento di Chimica Generale, Università di Pavia, Italy*

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ABSTRACT: The spatial and vertical distributions of six clay mineral assemblages were identified in the pelagic marls and marly limestones and in the Ammonitico Rosso of the Lower Jurassic in the Betic Cordilleras (SE Spain). The six assemblages contain varying proportions of illite, kaolinite, chlorite, smectite and mixed-layer illite-smectite. The clay mineralogy and the geochemical data, particularly for the rare earth elements, indicate that the sediments were deposited in a pelagic environment influenced by a nearby continent. The source areas were many and lithologically varied. Clay mineralogy and geochemistry are shown to be most useful in the reconstruction of the palaeogeography of a basin in a passive continental margin.

At the beginning of the Jurassic a considerable marine transgression took place in the Betic Cordilleras and a shallow carbonate platform developed with episodes of tidal-flat and, locally, open-sea facies. The environment is interpreted as being that of a very shallow carbonate platform, the varying textures of which reflect different energy regimes according to their position relative to the tidal level (García Hernández *et al.*, 1980). The deposition of sequences of the Ammonitico Rosso facies and thick marine series of marls and marly limestones show that pelagic conditions were established in the Subbetic Zone from the Domerian onwards.

In this paper we study the mineralogy and geochemistry of these facies with a view to understanding the palaeogeographic conditions that controlled their sedimentation, and also to locating the source-areas of these sediments.

GEOLOGICAL SETTING

The Betic Cordilleras, SE Spain, are part of the peri-Mediterranean Alpine orogenic belt (Fig. 1A). Two well-defined geological zones have been recognized within the chain, the External Zones and the Betic Zone (Internal Zone) (Fallot, 1948; Fontboté, 1970) (Fig. 1B). In the External Zones, composed of Mesozoic and Cenozoic rocks deposited on a continental margin along the southern border of the Meseta (Spanish Plain), Blumenthal (1927) and Fallot (op. cit.) distinguished two palaeogeographic realms, the northern Prebetic Zone and the southern Subbetic Zone. The former is characterized by shallow-water facies and the latter by a prevalence of pelagic facies. Three main

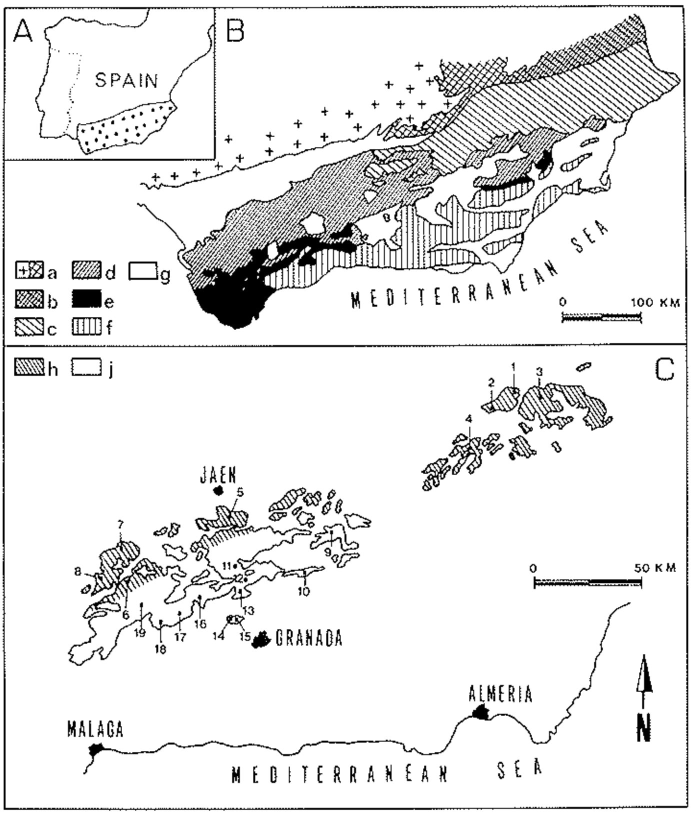

FIG. 1. Geological setting. A: location of the Betic Cordilleras. B: geological Complexes in the Betic Cordilleras. a: Iberian Plain and their cover; b: Iberian Domain; c: Prebetic Zone; d: Subbetic Zone and Penibetic; e: Flysch of Campo de Gibraltar; f: Internal Zones; g: Postorogenic basins. C: stratigraphic sequences studied. h: External Subbetic. 1: Majarazán (MAJ); 2: Guarrumbre (GU); 3: Fuente Vidriera (FV); 4: Huéscar (HU); 5: La Cerradura (LC); 6: Cañada del Hornillo (CH); 7: Pozo del Algar (PA); 8: Camarena-Lanchares-Zuheros (CLZ). j: Median Subbetic 9: Cerro Méndez (CM); 10: Zegrí norte (ZN); 11: Zegrí sur (ZS); 12: Iznalloz (IZ); 13: Colomera (CO); 14: Sierra Elvira (SE); 15: Toarcense Sierra Elvira (TSE); 16: Illora (I); 17: Sierra Pelada (SP); 18: Algarinejo (A); 19: Huétor Tajar (HT).

domains can be distinguished in the Subbetic Zone by their Mesozoic sequences from north to the south: the External, the Median and the Internal Subbetic. All the stratigraphic sequences described here belong to the Median and External Subbetic (Fig. 1C).

The macrofauna in the rhythmic alternations of marls and marly limestones consist fundamentally of ammonoids, although there are also some bivalves, gastropods, echinoids and brachiopods. Among the benthonic organisms foraminifera, ostracoda and radiolaria predominate. Packestone, wackestone and mudstone textures are common (Palomo, 1987). These lithofacies have been reported as being laid down in a marine-trough environment after the break up of the Liassic carbonate platform (Seyfried, 1979; García Hernández *et al.*, 1980).

The Ammonitico Rosso facies are made up of alternations of reddish marls and nodulous marly limestones and correspond to Aubouin's Ammonitico Rosso marneaux (1964). Braga *et al.* (1981) indicated that these facies are pelagic and were deposited at the bottom of the basin slopes.

Both types of lithofacies outcrop in the Zegrí Formation from the lower Domerian until the Aalenian, and very locally in the Carixian and Bajocian.

METHODS

Two hundred and thirty samples were studied in palaeontologically well-dated stratigraphical sequences, thus permitting accurate correlations between them. Thin sections of representative samples of the different lithofacies were prepared for petrographic and microfacies analyses.

The whole sample, clay and silt fractions were examined by X-ray diffraction (XRD) using a Philips PW 1710 X-ray diffractometer, and for quantitative analyses the intensity factors of Schultz (1964) and Barahona (1974) were used. The $MgCO_3$ content of the calcite was calculated by Harker & Tuttle's method (1955) and compared to the data of Liborio & Mottana (1973).

The crystal-chemical parameters of the phyllosilicates were determined by the methodology used by Palomo (1987) and Nieto *et al.* (1989). The basal spacings were measured using the MAX diffractometer program, which fixes the reflection maximum by measuring intensities at regular angular intervals for equal periods of time. The experimental conditions were: 0·005° 2θ step size between points, 1 s measuring time at each point, 40 points on each side of start angle. Quartz was used as an internal standard whenever present, and added if necessary. The basal spacing of the illite was measured from the 00,10 reflection, and that of the chlorite from the 004 reflection in oriented aggregates solvated with dimethylsulphoxide, or else from the 003 reflection in untreated samples. The crystallinity of the phyllosilicates was determined at the 002 reflection (IK, Kübler, 1968) for the illite, and the 003 reflection (width at half-maximum) for the chlorite using a scan speed of 2°/min, at time constant of 1, and a paper speed of 50 mm/°2θ.

The intensity ratios of the 002, 004 and 00,10 illite reflections and the 003, 004 and 005 chlorite reflections were measured on graphs obtained at a scan speed of 1°/min, a time constant of 2, and paper speeds of 10 mm/°2θ for illite, and 20 mm/°2θ for chlorite. The Fe content of the chlorite was calculated from the intensity ratios of the basal reflections, according to the method described by Nieto & Rodríguez Gallego (1983).

The morphological study of the clay minerals, the intergranular relationships of the various minerals and the distribution of the carbonate and silicate in the samples were carried out by scanning electron microscopy (SEM) using an ISI instrument equipped with EDX Kevex 5100-C.

The mica and chlorite micro-intergrowths in rocks from the presumed source area were analysed with a Camebax SX-50 automatic electron microprobe in the wavelength dispersion mode at an acceleration voltage of 20 kV, probe current of 5 nA, and electron beam diameter 0·5 μm. Albite (Na), orthoclase (K), wollastonite (Ca, Si), Al_2O_3, Fe_2O_3, periclase (Mg) and $MnTiO_3$ (Mn, Ti) were used as standards.

The chemical analyses of major and trace elements were carried out by X-ray fluorescence spectrometry (XRF) using a Philips PW 1410 spectrometer, using the methods proposed by Franzini *et al.* (1975) and by Leoni & Saitta (1976) to correct for matrix effects. The loss-on-ignition values were determined after ignition at 900°C. The rare-earth elements (REE) were quantified by instrumental neutron activation analysis using a 250 kW

Triga Mark II reactor. The usual procedure was to put 0·300 g of sample and rock standard (BCR-1 basalt, provided by USGS, and GS-N calco-alkaline granite, provided by ANTR, France) into polyethylene vials for irradiation. The elements analysed were La, Ce, Nd, Sm, Eu, Gd, Tb, Ho, Tm, Yb and Lu. The results were normalized by chondrites and NASC (North American Shale Composite). Induced radioactivity was measured with a Ge(Li) detector coupled to an analyser-computer for 5, 7, 10 and 50 days after the final irradiation.

MINERALOGY

Whole sample

The sequences belonging to the External and Median Subbetic contain varying quantities of calcite (C) + dolomite (D) (50–52%), clay minerals (CM) (38–39%) and quartz (Q) + feldspars (Fd) (9–12%). The average quantitative mineralogical variations, the standard deviations and values for the coefficient of variation (Palomo, 1987; Ortega-Huertas, 1990) reveal that all of these mineral groups vary much more widely in the Median Subbetic than in the External Subbetic. There are important mineralogical differences in specific areas and at specific times within the sedimentary basin (Fig. 2). Thus, during the lowermost Toarcian the quantitative mineralogy is most variable in both the External and Median Subbetic. This is particularly obvious in the latter zone where the sediments are notably heterogeneous from the middle Domerian onwards, whereas in the External Subbetic the

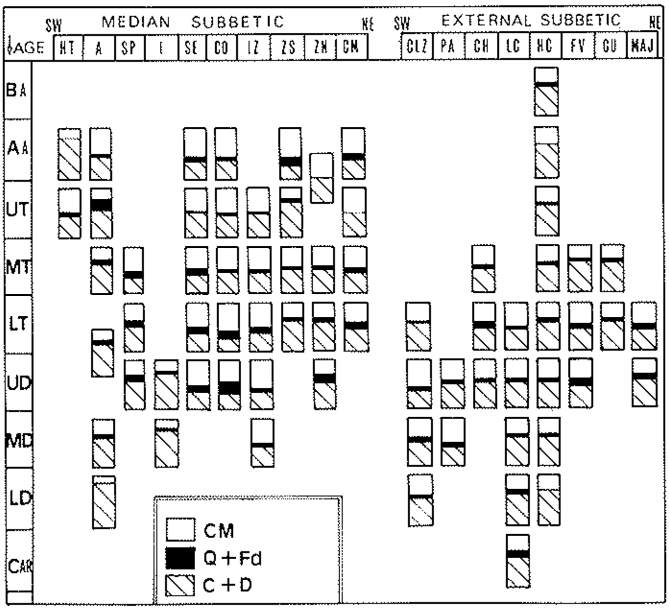

FIG. 2. Temporal and spatial variations of the whole mineralogy. BA: Bajocian; AA: Aalenian; UT: Upper Toarcian; MT: Middle Toarcian; LT: Lower Toarcian; UD: Upper Domerian; MD: Middle Domerian; LD: Lower Domerian; CAR: Carixian. CM: clay minerals; C: calcite; Q: quartz; D: dolomite; Fd: feldspars.

environment was more homogeneous with respect to both the quantity and nature of the deposits. Clay mineralogical analysis supports this hypothesis.

As mentioned above, carbonates form an important mineral phase in these lithofacies. SEM studies revealed occasional subhedral calcite crystals, and, in some samples from the upper Toarcian, dissolution voids can be distinguished indicating deep pelagic deposition.

The differences in the proportions of calcite found in the various sequences analysed are mainly due to the proportion of either carbonate-shelled or silicate-shelled organisms among the nannoplankton. In some sections, however, there is also an overall decrease in benthonic foraminifera towards the middle of the lower Toarcian and a faunal crisis at the Domerian-Toarcian boundary, characterized by euxinic conditions, the disappearance of benthonic organisms, and an abundance of juvenile forms (Braga *et al.*, 1982; Mira, 1987).

The calcite, which is a basic constituent of the carbonate phase, has an average $MgCO_3$ content of 1·80% mols. This corresponds to a non-magnesian calcite, typical of a pelagic marine environment, in which the carbonate contribution came essentially from foraminifera and coccolithaceae. These data agree with those of Schuber (1984) for the External Zones of the Betic Cordilleras at the Jurassic-Cretaceous boundary. Dolomite is scarce and irregularly distributed in sequences (<12%). It could have originated from the diagenetic replacement of calcite by a later exsolution process, as proposed by Müller *et al.* (1972). The mineralogical data and the presence of calcite particles within the dolomite rhomboids (Palomo, 1987) would seem to confirm this hypothesis.

Clay mineralogy

The clay minerals consist of illite (I), smectite (S), kaolinite (K), chlorite (Chl) and mixed-layer I-S. Quantitatively the average clay mineral values are similar in both the External and Median Subbetic; illite predominates (81% in External Subbetic and 75% in Median Subbetic), followed by smectite (11% in both formations) and minor quantities of the other minerals. Nevertheless, considerable differences do exist between cross sections, more particularly in the Median Subbetic than in the External Subbetic (Fig. 3).

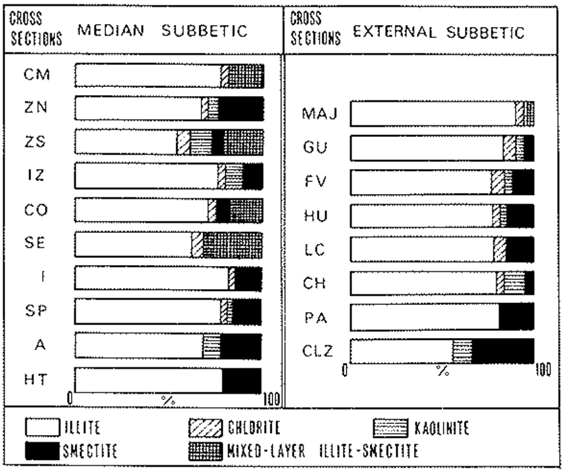

FIG. 3. Average clay mineral values in stratigraphical sequences.

As regards crystal chemistry, XRD studies revealed that the white micas are illites with a tendency towards phengites, particularly in the silt fraction. This agrees with the compositional tendencies obtained by applying Rey & Kübler's hypothesis (1983). Again there is little difference between the two geological realms as a whole, but changes are present between one section and another, and these must be attributed to the palaeogeographic situations of these sequences relative to the source areas. In the External Subbetic and Median Subbetic, the b_0 values for the illites vary between 8·997 and 9·017 Å, and between 9·007 and 9·021 Å in the clay fraction, while in the silt fraction they vary between 9·017 and 9·030 Å and between 9·007 and 9·037 Å, respectively. For the phengites this parameter is significantly consistent (the values of σ_x are $<0{\cdot}01$) for the sediments of the same age in both geological realms, suggesting that the source area changed little during the deposition of a particular sequence. On a wide scale it also suggests that both geological realms were supplied from the same source rocks. The basal spacing of the phengites supports this hypothesis, as significant differences exist between sequences but not within individual sections. In the External Subbetic the values range between 9·96 and 9·999 Å, and in the Median Subbetic between 9·97 and 9·99 Å.

Smectite is present in almost all the stratigraphic sequences, although in very variable proportions (Fig. 3). In some sequences it is one of the characteristic minerals of clay mineral association, its chemical composition falling between that of montmorillonite and beidellite, according to the results obtained using Greene-Kelly's treatment (1953). Chlorite is present in all the sequences studied, although in amounts <10% (Fig. 3), crystallinity ranging between 0·28° and 0·35° 2θ in the silt fraction, and 0·36° to 0·40° 2θ in the clay fraction. Kaolinite and mixed-layer I-S are scarce and irregularly distributed, tending to be concentrated in certain specific sequences in the basin (Fig. 3).

It is suggested that most of the clay minerals in these Jurassic argillaceous sediments are inherited from the source areas. Thus, the chlorite and part of the white micas are derived from the original rocks of the source area and the smectites and kaolinite from the soils developed upon them. The chlorite and phengite of the silt fraction reflect the general characteristics of the source area, although with slight modifications to some of their crystal-chemical parameters. For instance, the crystallinity index values of the mica range from 0·22 to 0·50° 2θ, compared to crystal sizes >1000 Å (López Munguira, 1987) in the source area micas. The same applies to the chlorites, which have a much smaller crystal size (400 Å) in the sediments than in the low-grade and very low-grade metamorphic rocks of the source area. The coarse-grained phengites, with tabular habits and a K/Fe ratio between 1·58 (External Subbetic) and 0·90 (Median Subbetic), are typically inherited from these types of rocks. The high Fe content may be due to the existence of submicroscopic intergrowths of chlorite within the mica in the source area rocks and indeed our studies of the rocks from various lithological complexes of the Spanish Plain support this hypothesis. Thus, for example, the most common crystals are mica and chlorite micro-intergrowths of 1–2 μm (Fig. 4), occurring within crystals between 8 and 30 μm in size. Analyses revealed the following representative compositions:

Chlorite (chamosite): $(Si_{3\cdot23}Al_{0\cdot77})(Al_{1\cdot92}Ti_{0\cdot01}Fe_{2\cdot24}Mg_{1\cdot12}Mn_{0\cdot02})O_{10}(OH)_8$

Mica: $(Si_{3\cdot17}Al_{0\cdot83})(Al_{1\cdot91}Fe_{0\cdot06}Ti_{0\cdot02}Mg_{0\cdot04}(K_{0\cdot67}Na_{0\cdot16})O_{10}(OH)_2$

Analogous conclusions were reported by Franceschelli *et al.* (1986) in low-grade metamorphic metapelites and by Morad (1986) in argillaceous sediments from very low-grade metamorphic rocks. In addition to this general process of inheritance, which affected

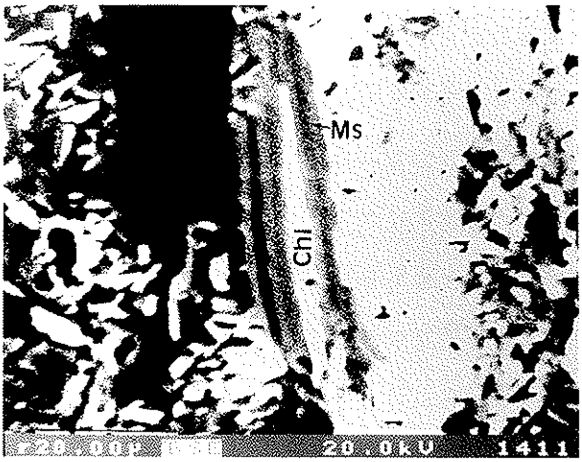

FIG. 4. Back-scattered electron micrograph showing a chlorite (Chl)- muscovite (Ms) intergrowth. See composition in text.

mainly the silt fraction and to some extent the clay fraction, there is a diagenetic smectite-illite transformation. This is evident in various sequences in the Median Subbetic (Palomo *et al.*, 1985). It is particularly clear in the SE cross-section where, from top to bottom, there is an increase in the crystallinity index (IK) (1·30° 2θ to 0·40° 2θ) and in the proportion of illite (5 to 88%) and also a decrease in the smectite and mixed-layer I-S content (90 to 9%). SEM reveals that the illite presents a plate and straight blade habits, and a K/Fe ratio whose mean values range from 2·68 (External Subbetic) to 1·28 (Median Subbetic).

GEOCHEMISTRY

The XRF data are given in Tables 1 and 2.

The trace elements are concentrated mainly in the clays, with the exception of Sr, Y and Zr; Sr is associated with the carbonate phase, and Y and Zr are normally transported with the accessory minerals (especially zircon). This high density also results in their being partially excluded from the <2 μm fraction during granulometric separation.

The K/Rb ratio (Fig. 5A) in the clay fraction of the marls and marly limestones is higher in the External Subbetic than in the Median Subbetic. Rubidium is related fundamentally to illite but also to some extent to smectite. Therefore the more pelagic the depositional environment, the lower is the K/Rb ratio. In the samples studied, Zr varies between 100 and 125 p.p.m., both in the whole sample and the clay formation, but Rb is concentrated in the clay fraction in quantities ranging between 125 and 150 p.p.m. in the Median Subbetic, and between 125 and 175 p.p.m. in the External Subbetic. Titanium and Nb are concentrated in the clay fraction, while Zr and Y are distributed evenly in both the whole sample and the clay fraction (Fig. 5B). The content of these four elements is closely dependent upon the clay mineralogy, being related above all to those associations richer in illite and smectite. Thus, in some samples we have identified exsolutions of anatase upon illite (Palomo, 1987) and by analytical electron microscopy show the widespread existence of Ti in both illites and smectites.

The total REE content, the Ce/Ce* ratio and the proportion of carbonates are shown in

TABLE 1. Geochemical analyses of the bulk samples.

		%											p.p.m.								
Sample	L	SiO_2	Al_2O_3	CaO	MnO	MgO	Na_2O	K_2O	TiO_2	Fe_2O_3	P_2O_5	LOI	Ba	Rb	Sr	Y	Zr	Nb	V	Cr	Ni
External Subbetic																					
MAJ-5	(1)	34·12	10·28	22·33	0·04	2·07	0·05	3·26	0·62	3·90	0·04	23·28	213	92	349	18	118	14	85	66	33
GU-8	(1)	17·76	5·57	35·99	0·03	1·63	0·00	1·22	0·30	1·67	0·03	35·61	122	42	383	11	50	7	40	42	21
GU-4	(1)	20·94	6·13	33·59	0·05	1·72	0·00	1·37	0·35	2·42	0·03	33·41	132	48	360	14	67	9	42	39	24
FV-19	(1)	19·52	6·17	34·48	0·04	1·78	0·00	1·28	0·34	2·01	0·03	34·34	260	47	360	14	65	8	43	36	23
FV-1	(1)	30·73	8·61	26·20	0·04	2·08	0·14	2·31	0·52	3·06	0·05	26·27	211	82	508	16	103	10	70	59	35
HU-11	(1)	26·99	6·51	29·96	0·04	2·01	0·00	1·74	0·38	3·05	0·02	29·30	154	61	487	13	81	9	49	46	30
CH-10	(1)	36·34	10·95	21·42	0·03	2·07	0·10	3·00	0·72	3·25	0·06	22·05	230	87	163	20	145	16	84	67	37
CH-1	(1)	23·82	7·88	30·24	0·04	1·44	0·00	2·10	0·48	2·90	0·04	31·07	167	66	212	17	103	12	55	48	31
PA-10	(1)	24·14	7·12	30·96	0·03	1·81	0·00	2·04	0·44	2·52	0·04	30·90	156	65	350	15	103	10	63	50	35
PA-1	(1)	28·58	8·81	27·33	0·03	2·07	0·00	2·38	0·53	2·64	0·03	27·60	165	79	215	16	120	11	73	62	46
CLZ-5	(1)	26·71	9·08	27·87	0·04	1·62	0·00	1·71	0·53	3·31	0·06	29·07	178	73	159	15	112	13	61	57	32
CLZ-2	(1)	23·92	7·05	31·29	0·03	1·79	0·00	1·38	0·44	2·73	0·04	31·34	147	66	240	19	117	9	55	51	30
Median Subbetic																					
ZN-14	(2)	17·49	5·38	35·42	0·11	1·35	0·00	1·12	0·30	2·75	0·05	36·03	105	41	175	15	54	8	37	30	24
ZN-7	(1)	26·88	8·95	25·98	0·06	3·81	0·04	1·82	0·50	4·13	0·07	27·75	193	69	469	18	92	13	68	52	30
IZ-19	(2)	14·85	4·64	37·74	0·10	0·98	0·00	1·04	0·27	2·28	0·04	38·07	124	33	105	12	46	9	32	25	39
IZ-1	(1)	19·19	5·68	35·56	0·04	1·45	0·00	1·31	0·33	1·71	0·03	34·71	104	55	191	11	69	8	35	42	23
TSE-13	(3)	36·43	10·42	20·87	0·06	3·03	0·07	2·59	0·62	4·18	0·05	21·70	166	83	353	18	114	15	81	60	37
TSE-8	(3)	29·59	8·42	26·55	0·06	2·52	0·04	1·99	0·51	3·39	0·06	26·87	150	66	378	19	99	11	65	48	29
TSE-3	(3)	40·67	11·75	16·97	0·05	3·04	0·08	2·96	0·71	5·24	0·07	18·45	207	102	334	23	148	18	95	72	47
I-5	(1)	14·41	3·32	40·62	0·05	1·26	0·00	0·84	0·20	1·29	0·03	37·98	82	27	383	6	29	6	24	28	15
SP-10	(2)	22·60	7·12	31·58	0·09	1·38	0·00	1·48	0·41	3·04	0·04	32·25	158	56	201	16	84	8	56	38	25
SP-5	(1)	30·80	5·80	30·59	0·02	1·30	0·00	1·25	0·35	1·74	0·02	28·13	120	49	666	11	77	8	42	55	23
SP-1	(1)	37·31	6·75	26·11	0·03	1·37	0·00	1·66	0·42	2·25	0·04	24·05	141	56	433	13	95	11	54	64	30
A-16	(1)	33·23	8·90	24·71	0·05	2·28	0·03	1·90	0·55	3·52	0·07	24·78	188	82	382	23	96	12	71	63	40
A-8	(1)	20·63	5·58	35·54	0·03	0·85	0·00	0·92	0·34	1·81	0·03	34·27	118	38	355	11	72	9	33	37	19
HT-2	(2)	32·52	8·25	25·11	0·15	2·02	0·02	1·96	0·46	4·13	0·10	25·29	158	73	193	22	68	12	56	52	57

L = Lithology.
(1) Marls and marly limestones.
(2) Ammonítico Rosso.
(3) Turbiditic levels.

TABLE 2. Geochemical analyses of the clay fractions.

		%							p.p.m.								
Sample	L	SiO_2	Al_2O_3	MnO	K_2O	TiO_2	Fe_2O_3	LOI	Ba	Rb	Sr	Y	Zr	Nb	V	Cr	Ni
External Subbetic																	
MAJ-5	(1)	50·79	19·88	0·03	5·94	0·89	8·64	7·86	255	187	159	18	128	21	189	134	57
MAJ-3	(1)	52·63	19·14	0·02	5·74	0·86	7·62	7·38	235	183	111	18	123	20	172	126	65
MAJ-1	(1)	53·08	18·05	0·03	5·45	0·84	7·04	7·85	246	167	118	14	109	19	150	123	61
GU-11	(1)	48·57	18·72	0·03	4·42	0·97	7·13	10·21	266	150	170	18	118	22	164	119	65
GU-8	(1)	49·97	19·94	0·02	4·58	0·87	6·02	9·27	249	151	257	17	108	19	174	120	50
GU-4	(1)	49·62	18·23	0·02	4·28	0·96	7·35	9·77	244	134	234	16	114	22	149	112	50
FV-21	(1)	52·69	18·86	0·03	3·67	0·96	7·67	7·50	250	125	138	14	99	20	181	121	53
FV-19	(1)	50·49	19·80	0·02	4·33	0·99	7·56	8·22	290	148	205	19	120	22	168	122	55
FV-14	(1)	52·22	18·50	0·02	4·95	0·86	9·02	7·35	278	135	141	13	85	18	166	127	76
Fv-1	(1)	52·91	18·91	0·03	5·41	0·87	7·07	7·16	271	175	155	15	113	20	169	134	60
HU-23	(1)	71·82	11·07	0·02	2·33	0·66	5·04	3·42	182	101	58	14	70	17	98	71	95
HU-18	(2)	50·21	17·18	0·03	4·31	0·77	11·17	7·83	233	151	66	15	97	17	134	107	67
HU-14	(1)	54·03	20·32	0·03	4·67	1·21	6·59	6·94	286	157	190	18	160	27	158	119	47
HU11	(1)	69·49	11·72	0·03	3·28	1·03	4·20	4·43	367	109	106	24	188	24	85	86	50
HU-5	(1)	58·56	16·43	0·03	4·53	0·78	6·26	5·55	281	161	114	12	105	18	141	116	64
CH-10	(1)	43·90	17·07	0·03	4·28	0·84	6·11	13·26	221	144	148	19	119	19	160	117	55
CH-7	(1)	42·33	18·81	0·03	4·51	0·88	5·89	14·26	237	149	214	21	132	20	165	121	51
CH-5	(1)	44·49	18·69	0·03	4·38	0·84	7·71	12·75	236	151	191	22	132	20	160	132	63
CH-3	(1)	41·34	17·87	0·04	4·54	0·95	8·22	13·94	250	148	187	20	132	21	155	116	64
CH-1	(1)	43·32	18·02	0·04	4·66	1·04	7·79	12·88	253	153	170	21	144	24	151	116	66
PA-10	(1)	42·87	15·95	0·04	4·64	0·81	7·75	14·81	243	160	142	15	103	18	165	129	64
PA-1	(1)	45·04	17·44	0·03	4·99	0·97	6·09	12·88	262	171	117	16	124	22	171	137	87
CLZ-5	(1)	46·57	20·60	0·05	3·54	1·20	8·31	11·08	307	172	198	27	170	29	164	128	62
CLZ-2	(1)	40·44	14·67	0·03	2·83	0·67	6·53	16·71	190	145	161	19	110	17	127	106	52

TABLE 2—*continued.*

		%							p.p.m.								
Sample	L	SiO_2	Al_2O_3	MnO	K_2O	TiO_2	Fe_2O_3	LOI	Ba	Rb	Sr	Y	Zr	Nb	V	Cr	Ni
Median Subbetic																	
ZN-14	(2)	43·60	15·33	0·06	3·47	0·55	9·28	11·84	167	126	90	10	75	17	110	86	54
ZN-7	(1)	44·16	17·55	0·04	3·95	0·69	8·62	12·25	226	137	239	17	103	19	145	110	49
ZN-4	(1)	58·24	15·47	0·02	3·68	0·68	4·75	7·53	218	154	123	5	67	18	127	152	51
ZN-3	(1)	58·37	15·47	0·02	3·68	0·68	4·73	7·48	213	155	123	5	68	16	129	155	50
IZ-19	(2)	46·01	18·48	0·09	4·13	0·90	11·68	9·95	238	140	128	16	118	19	143	108	82
IZ-14	(2)	44·95	20·24	0·06	4·13	1·01	12·23	10·48	274	151	155	20	133	22	205	127	60
IZ-9	(2)	53·80	20·96	0·03	5·00	1·31	7·55	5·64	299	173	176	23	176	31	168	123	67
IZ-4	(1)	52·24	19·62	0·02	4·90	1·05	7·72	7·31	254	193	98	19	143	24	137	134	67
IZ-1	(1)	50·88	18·98	0·03	4·87	1·29	8·60	8·08	291	214	80	16	154	32	151	161	63
I-5	(1)	50·22	13·66	0·02	3·70	0·65	5·88	10·94	175	151	110	14	92	18	119	129	44
I-4	(1)	50·81	12·49	0·03	3·24	0·68	5·09	11·48	187	127	102	11	84	18	117	125	50
I-3	(1)	47·71	17·26	0·04	4·83	0·98	10·41	9·00	224	159	68	11	109	22	216	112	124
SP-10	(2)	47·84	20·25	0·04	4·39	1·15	10·21	9·07	294	162	246	25	158	26	194	125	59
SP-5	(1)	66·06	13·55	0·01	3·11	0·72	5·06	5·87	186	131	108	11	90	20	118	134	49
SP-3	(1)	66·38	13·42	0·02	3·17	0·73	4·33	4·56	189	129	111	10	83	19	122	156	49
SP-1	(1)	67·11	12·97	0·02	3·39	0·69	5·27	4·04	201	126	101	15	81	18	125	139	50
A-16	(1)	50·10	15·56	0·02	3·41	0·76	6·80	9·05	214	146	153	23	108	18	133	112	58
A-12	(1)	42·42	16·28	0·04	3·06	0·80	5·62	15·39	256	126	298	24	108	19	128	99	52
A-10	(1)	46·18	17·35	0·02	2·75	0·77	6·00	12·37	195	119	321	17	98	20	113	92	42
A-8	(1)	42·86	14·85	0·02	2·62	0·69	5·33	16·24	201	127	191	12	87	18	100	95	38
A-4	(1)	59·39	15·72	0·02	3·06	0·74	5·19	7·34	215	138	262	13	90	19	128	147	72
A-3	(1)	63·21	13·74	0·02	2·59	0·66	4·53	7·00	191	113	149	10	69	17	106	177	69
HT-2	(2)	52·36	17·08	0·03	4·17	0·71	8·35	7·85	223	172	88	20	112	17	136	104	111

L = Lithology.
(1) Marls and marly limestones.
(2) Ammonítico Rosso.

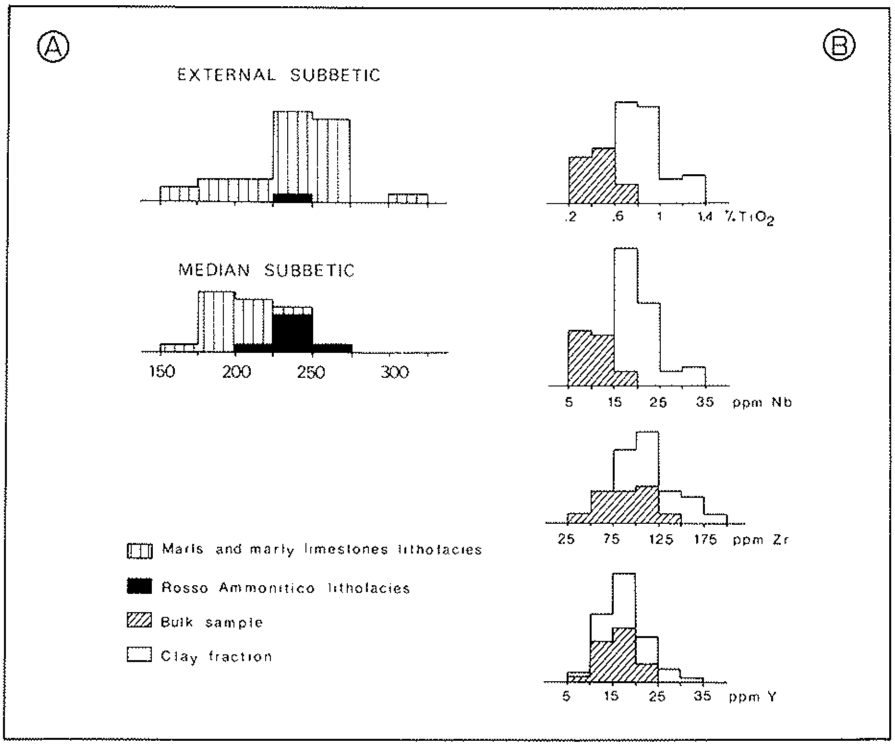

FIG. 5. A: K/Rb ratio in clay fraction. B: chemical composition of whole samples and clay fractions.

Table 3. (Ce* corresponds to the concentration obtained by extrapolation between La and Nd, as described by Courtois & Hoffert, 1977). The REE measurements were made on the whole sample and thus it was necessary to determine precisely their relationship to individual mineral phases. The correlation coefficient between the REE and carbonates is negative ($r = -0{\cdot}75$) and thus the REE must be related to the aluminosilicate phase and/or the hydroxides. The margin of variation in the REE content of the marls and marly limestones is shown in Fig. 6. While the overall values are higher in the External Subbetic, the evolution of the chondrite-normalized REE curves is analogous in both geological realms. In the TSE sequence (Median Subbetic) the higher REE content may well be attributable to the turbiditic origin of the sediments (Fig. 6). The variations in the Ammonitico Rosso lithofacies are similar to those in the marls and marly limestones of the Median Subbetic.

The average Ce/Ce* values in relation to the age of the sediments are summarized in Table 4. In general the relationship is higher in the External Subbetic, implying a greater detrital influence in this part of the basin. The high values found locally in some sequences of the Median Subbetic are due to the erosion of emerged areas within the basin at certain periods during its development. An example of this is the paleokarst adjacent to the A and SP stratigraphical sequences (cf. Vera *et al.*, 1989).

TABLE 3. Total REE content, Ce/Ce* ratio (see text) and %$CaCO_3$ in whole samples.

External Subbetic					Median Subbetic				
Lithology	Sample	Total REE p.p.m.	Ce/Ce*	%$CaCO_3$	Lithology	Sample	Total REE p.p.m.	Ce/Ce*	%$CaCO_3$
1	MAJ-3	114	0·81	34	1	ZN-10	112	0·67	50
1	GU-11	84	0·72	56	1	ZN-3	93	0·63	49
1	GU-1	206	0·79	55	2	IZ-14	111	0·62	50
1	FV-10	76	0·84	44	1	IZ-4	111	0·69	30
2	HU-18	100	0·72	6	3	TSE-12	117	0·87	49
1	HU-5	78	0·74	64	3	TSE-10	110	0·92	70
1	CH-5	134	0·80	44	3	TSE-3	144	0·85	29
1	PA-7	84	0·63	47	3	TSE-1	135	0·90	29
1	CLZ-6	130	0·62	56	2	I-3	57	0·87	70
1	CLZ-1	62	0·53	55	1	SP-7	82	0·59	68
					1	SP-3	69	0·86	52
					1	A-12	71	0·72	60
					1	A-4	72	0·77	64
					2	HT-3	127	0·68	35

1: Marls and marly limestones. 2: Ammonítico Rosso. 3: Turbiditic levels.

PALAEOGEOGRAPHY AND SOURCE AREAS

In these Jurassic marl and marly limestone and Ammonitico Rosso lithofacies the following clay mineral associations were identified by XRD (Palomo *et al.*, 1985; Ortega-Huertas *et al.*, 1985; Palomo, 1987):

Association A = I, Chl, K;
Association B = I, Chl;
Association C = I, Chl, S;
Association D = I, Chl, I-S;
Association E = I, S;
Association F = I, K, S.

We consider Association A to be related to sediments in which the influence of emerged areas is evident, in agreement with various authors' interpretations (López Aguayo & Caballero, 1973; Singer, 1984; Andeol, 1985). The clay minerals of Association B are characteristic of Triassic detrital facies in the Betic Cordilleras and are analogous to the Triassic Keuper at the border of the Spanish Meseta (Palomo, 1987). This mineral association has also been described by Andeol (1985) as typical of a distal continental domain in other areas of the Betic Cordillera. The rest of the clay mineral associations represent pelagic conditions with the exception of Association F, which is restricted to two stratigraphic sequences and may be interpreted as reflecting climatic changes and substantial karstification and pedogenesis in the source area (Vera *et al.*, 1989).

The spatial and temporal distribution of these clay mineral associations is represented in Fig. 7. The sedimentation in the External Subbetic was initially influenced by emerged areas but later the sequences became more pelagic. The depositional basin in the Median

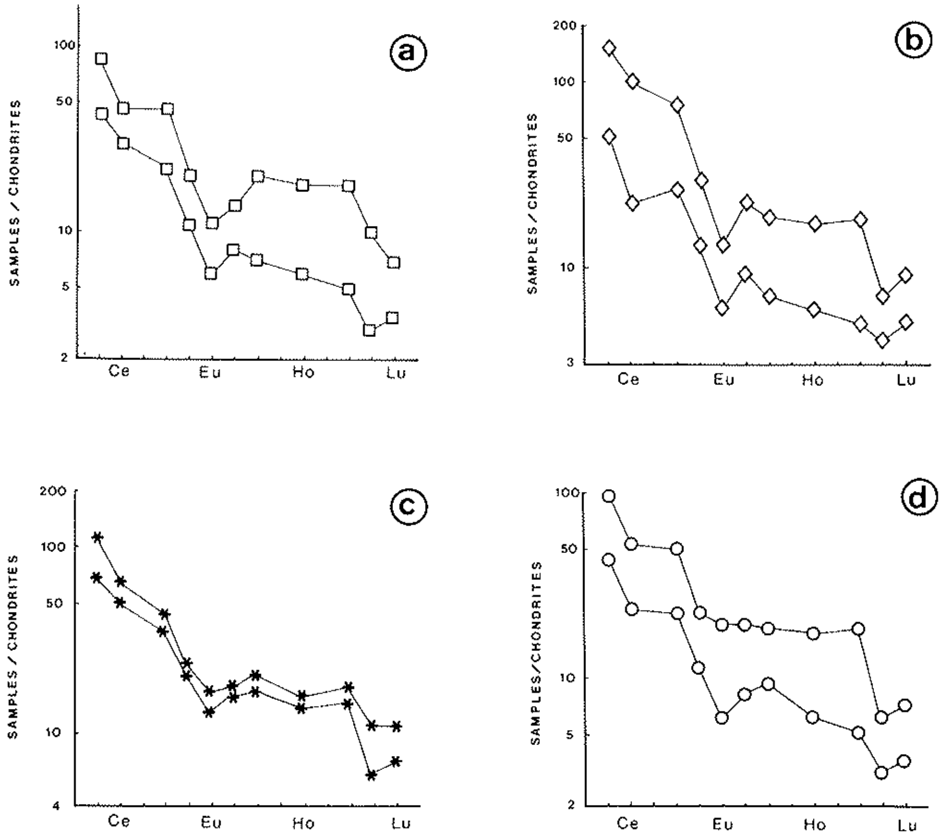

FIG. 6. Chondrite-normalized REE distribution. The curves correspond to the samples with maximum and minimum values. a: marls and marly limestones, Median Subbetic; b: marls and marly limestones, External Subbetic; c: Toarcian turbiditic sequence (TSE); d: Ammonitico Rosso lithofacies.

TABLE 4. Ce/Ce* values (see text) in relation to the age of the sediments.

Age	Median Subbetic Ce/Ce*	External Subbetic Ce/Ce*
Upper Toarcian–Aalenian	0·68	0·72
Middle Toarcian	0·65 (0·89)	0·79
Lower Toarcian	0·59	0·74
Upper Domerian	0·73	n.d.
Middle Domerian	0·78	0·69
Lower Domerian	n.d.	0·53

n.d. not determined.
( ) includes turbiditic levels.

Subbetic was less homogeneous. In general, the environment was one of a pelagic marine trough with several swells forming at different times and at different locations (see for example, cross-sections IZ, SP). This idea is also supported by the uniform V and Cr content of the marls and marly limestones in the External Subbetic and their great variability in the Median Subbetic (Table 2).

As a parameter of continental influence (Courtois & Hoffert, 1977) the spatial representation of the Ce/Ce* ratio also indicates a greater influx of detrital materials in the eastern sector of the External Subbetic, while the western sector was clearly a deep pelagic environment (Fig. 8A). The high Ce/Ce* value (0·80) in the CH sequence is due to the fact that this sample is rich in illite, chlorite, kaolinite and feldspars and relatively poor in $CaCO_3$ (Table 3). In the Median Subbetic the environment is more complex. Except for the TSE turbiditic sequence, the environment was mainly deep pelagic but with some areas of considerable continental influence, such as the A and SP sequences, in which the Ce/Ce* values are 0·74 and 0·73, respectively. Vera *et al.* (1989) reached similar conclusions on the basis of stratigraphical evidence. The 0·87 value in the I sequence is due mainly to the high proportion of iron and aluminium oxides and hydroxides, typical of Ammonitico Rosso lithofacies.

The clay mineral associations and the values of Ce/Ce* ratio indicate that, on the whole, in both the External Subbetic and Median Subbetic all the stratigraphic sequences are transgressive towards the top. Nevertheless, the high detrital content (D index) of the sediments in the Subbetic basin shows that there was a significant and continuous continental influence throughout the Jurassic (Fig. 8B). Similarly, the Zr/Rb ratio is a good indicator of the relative proportions of detrital and non-detrital fractions in the sediments; as carbonates hardly contain either of these elements, their proportions are related basically to the clay minerals, particularly those in which illite is most abundant. The Zr/Rb ratio in the External Subbetic is 1·50 and 1·40 in the Median Subbetic. Following the opinion of Taylor (1965) that this ratio diminishes towards the centre of the basin, and given that the Zr/Rb ratio drops to 1·33, it would appear that the Median Subbetic was most pelagic at the end of the Domerian.

An analysis of the relationship between the lithofacies and clay-mineral associations indicates that all of the lithofacies may have been deposited in different palaeogeographical conditions. Thus, for example, the Ammonitico Rosso was laid down both in troughs (ZS and CO sections) and on swells (IZ and SP sections). Furthermore, there is no direct relationship between the lithologies and the clay mineral associations, and indeed the marls and marly limestones are related to them all.

In summary, the lithofacies do not correspond to those of a homogeneous sedimentary environment but were deposited under different conditions according to their geographical situation in the basin, and also to the geological period in which they were laid down.

In trying to determine the source area of these Jurassic sediments, we have taken into account the distribution of the clay mineral associations, and the crystal-chemical criteria of Nieto *et al.* (1989), namely the b_0 parameter, crystal size and basal spacing of the micas, and the Fe content and basal spacing of the chlorites. This last parameter is considered to be of special interest as it is affected neither by erosion nor weathering, and variation depends directly upon the lithology of the source areas (Liebling & Scherp, 1980; Nieto & Rodríguez Gallego, 1983). These crystal-chemical parameters indicate that the Jurassic sediments in the Subbetic Zone are derived from several source areas. Following the interpretation of Stephens *et al.* (1975), the Al/Ti ratio of between 14 and 24 in our samples suggests a

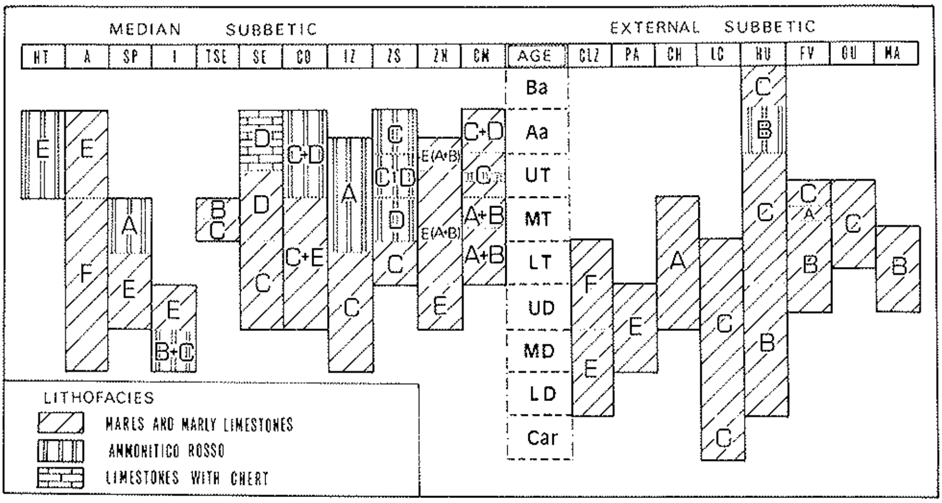

FIG. 7. Areal and vertical distribution of clay mineral assemblages and their relation to the lithology. A: I, Chl, K; B: I, Chl; C: I, Chl, S; D: I, Chl, I-S; E: I, S; F: I, K, S. (Legend for ages given with Fig. 2.)

continental source area rather than an oceanic one, which would give values closer to 5. The crystal-chemical data given earlier in this paper, together with the mean value of the chlorite basal spacing (14·10 Å), the Fe content in the chlorites (1·70 to 2·22 atoms per formula unit) and the widespread presence of coarse grained phengites, all indicate that the rocks of the Spanish Meseta, north of the Subbetic Zone, were the principal supplier of sediments to this basin. In particular, the very low-grade to low-grade metamorphic Precambrian greywacke schists from the Central Iberian Zone have the same crystal-chemical values as those of the sediments in question (Palomo, 1987; López Munguira, 1987). This source area must have been active throughout the depositional changes during this period, and the differences between stratigraphic sequences bear witness to the diverse lithology of the source area itself.

The local presence of the clay mineral association B in the easternmost area of the External Subbetic indicates that there was some influence from the Prebetic Triassic Keuper, which emerged during the Mesozoic (cf. García Hernández *et al.*, 1980).

Similarly, the significant quantities of kaolinite (clay mineral association A) or the presence of clay mineral association F show that some Ammonitico Rosso and marl and marly limestone deposits could well have been affected by locally emerged areas. Vera *et al.* (1989) proposed the existence of a clear relationship between the kaolinite and smectite (clay mineral association F) content of the Middle Domerian to Middle Toarcian sediments of the A sequence, and the emergence at the same time of a stratigraphic sequence some 10 km away. Here, karstification and pedogenetic processes took place on islands within the passive continental margin and these in turn affected the clay mineral composition of the nearby pelagic deposits. This hypothesis is also supported by geochemical data. For example, in samples from the emerged periods the Ce/Ce*, Rb/Zr and D index values are 0·77, 1·89 and 0·70, respectively, against 0·72, 1·17 and 0·60 in samples from the upper Toarcian-Aalenian, which are clearly pelagic in character. The influence of a local source area is also clear in the SP sequence, situated to the east of the A sequence, where the

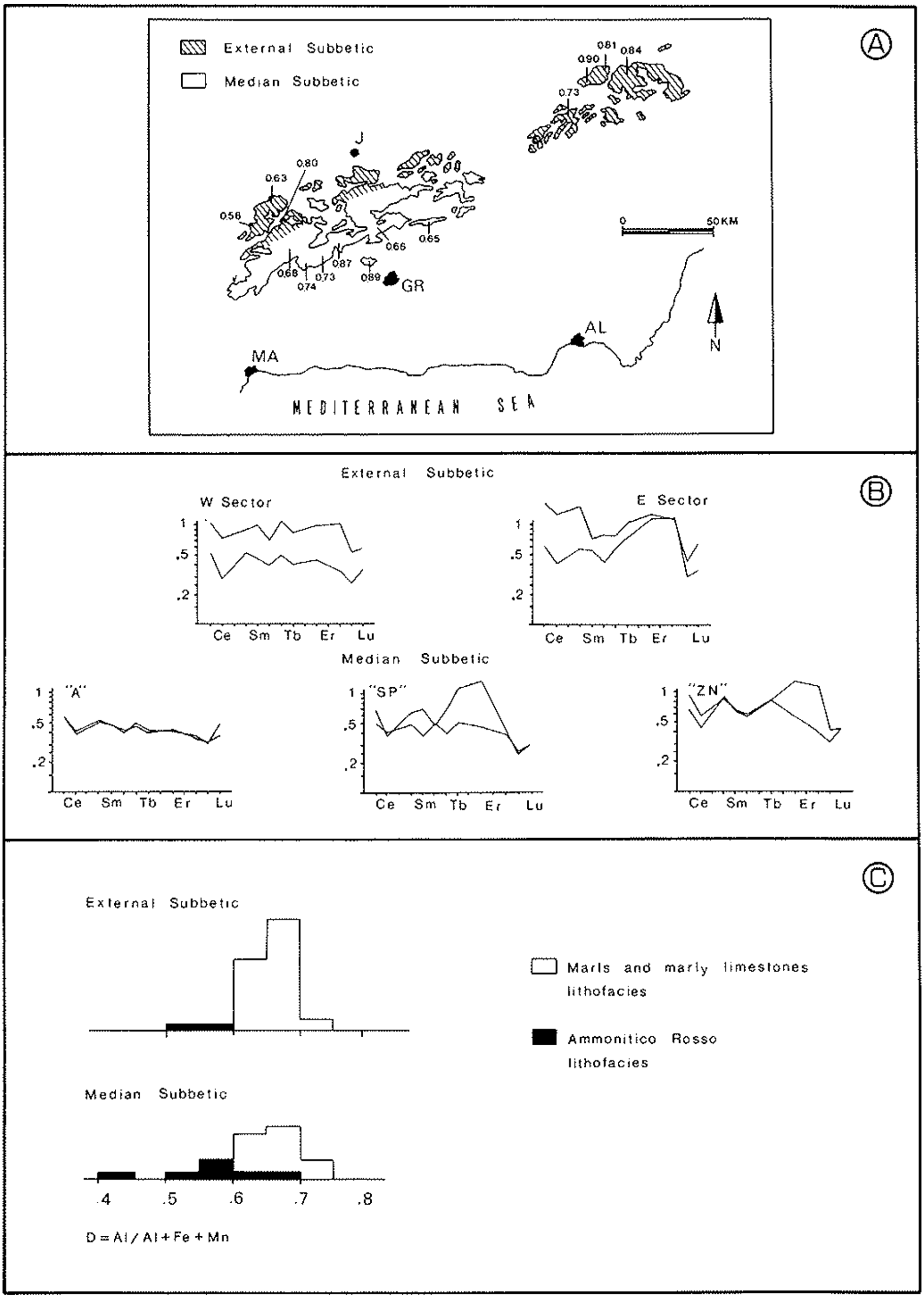

FIG. 8. A: distribution of Ce/Ce* ratio in the stratigraphical sequences studied. (Caption as Fig. 1). B: NASC-normalized REE distribution: range of values. "A": Algarinejo cross-section; "SP": Sierra Pelada cross-section; "ZN": Zegrí Norte. C: values of detrital index (D).

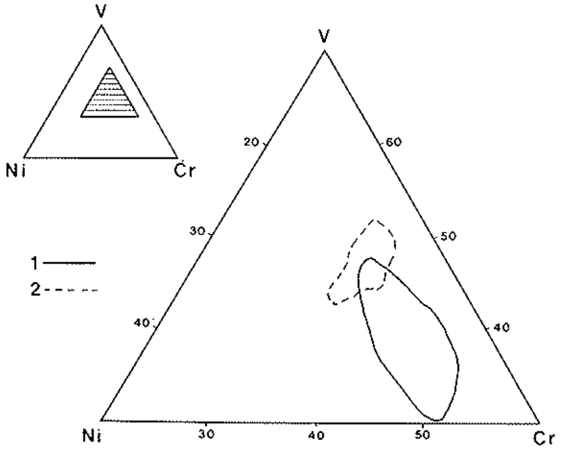

FIG. 9. Content of V, Ni and Cr in the marl and marly limestone lithofacies. 1: Median Subbetic; 2: External Subbetic.

Ce/Ce* and SiO_2/Al_2O_3 ratios reach maxima of 0·86 and 5 in the Upper Domerian, compared to 0·59 and 2·30 in the Middle Toarcian.

The relative quantities of V, Cr and Ni, elements usually associated with a detrital origin, point to similar conclusions in other areas of the Median Subbetic. In fact the most frequent V/Cr values range from 1·15 to 1·50 in the External Subbetic, and from 0·75 to 1·00 in the Median Subbetic where the general tendency is towards an enrichment in Cr (Fig. 9). All the samples richest in Cr in the Median Subbetic correspond to the Middle Domerian-Lower Toarcian, while those from the Middle Toarcian-Aalenian are similar in Cr content to the External Subbetic. This leads us to believe that in the Median Subbetic there was a considerable contribution of material from soils developed upon limestones in adjacent emerged areas, (stratigraphic sequences A, SP, ZN, IZ) up until the Lower Toarcian.

In other areas of the Median Subbetic (TSE sequence), the high Fe content (3·30–4·65 atoms per formula) and basal spacings (14·15–14·16 Å) of the chlorites in a turbiditic sequence suggest that they derived from the Lower Liassic carbonate platform. Similar values have been reported by Whittle (1986) for other sedimentary chlorites.

The variability of the basal spacings, together with the heterogeneous chemical composition of the smectites in these Jurassic sediments, support the idea of the local influence of diverse types of rocks and soils in the sedimentary environment.

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