MESSINIAN MARGINAL-MARINE AND CONTINENTAL FACIES AND THEIR STRATIGRAPHY IN THE EASTERN ALMERIA PROVINCE (SE SPAIN)

by

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to Claude, family and friends

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SAMENVATTING

De stratigrafie, micropaleontologie en geochemie van Neogene sedimenten van de Oostelijke Almeria Provincie ('OAP', Zuidoost Spanje) leveren nieuwe data met betrekking tot de evolutie van het lokale, regionale (mediterrane) and globale milieu vanuit een gebied wat goed gesitueerd is voor de bestudering van hun interactie. De voorliggende studie is geconcentreerd op het Messinien interval, wat de grens tussen de Miocene en Pliocene tijdvakken markeert en wat gekenmerkt wordt door de zog. 'Saliniteitscrisis' van het Mediterrane Bekken, waar zich tijdelijk een scala aan 'randbekken' en continentale omstandigheden ontwikkelen. De sedimenten uit het meer centrale deel van het Noordelijke Nijar of Carboneras Bekken en enkele aanvullende secties van kritische intervallen uit de aangrenzende Agua Amarga, Sorbas en Vera bekkens zijn in detail bestudeerd. Ter vergelijking zijn verkenningsstudies gemaakt van de afzettingen van de bekkenranden van de OAP en van enkele secties in de meer externe delen van de Boog van Gibraltar en het centrale deel van het Mediterrane Bekken.

De Oostelijke Almeria Provincie ligt op de westrand van het Mediterrane Bekken, relatief dicht bij de Straat van Gibraltar die tegenwoordig de enige verbinding vormt tussen de Middellandse Zee en de open ocaan. Gedurende het Mioceen bestonden er verschillende verbindingen in dit 'Mediterrane Ingangs Gebied' (MIG), dat gekenmerkt wordt door de aanwezigheid van het boogvormige Betisch-Rif-Kabylen orogeen of Boog van Gibraltar van zuid Spanje en het noordelijkst gedeelte van de Maghreb. De paleogeografische constitutie van dit 'Gibraltar Boog Drempel Systeem' (GBDS) is een belangrijke factor in de effectiviteit van de water-uitwisseling tussen Middellandse Zee en Atlantische Oceaan.

Binnen het Neogeen van de Oostelijke Almeria Provincie zijn een aantal opeenvolgende facies eenheden onderscheiden op basis van consitente verticale veranderingen in microfossiel inhoud en primaire en diagenetische lithologische kenmerken in de bekken centra en veranderingen in lithologische kenmerken, vaak gemarkeerd door erosie oppervlakken, aan hun randen. Deze facies eenheden zijn geinterpreteeerd in termen van opeenvolgende 'milieu stadia' van het gebied, terwijl hun grenzen zijn genummerd als 'events'. Stadia en\events zijn in de eerste plaats gedateerd door middel van planktonische foraminiferen biostratigrafie. De milieu ontwikkeling over een langere periode kan het best bestudeerd worden door vergelijking van de facieseenheden die de relatief stabiele perioden van hoog zeeniveau van de opeenvolgende afzettings cycli vertegenwoordigen.

De successievelijke dominantie van diep mariene afzetting gedurende het Mioceen, van ondiep mariene afzetting gedurende het Plioceen en van continentale afzetting gedurende het Kwartair, wordt onderbroken door kortere 'regressieve episoden', gekenmerkt door facies die restrictie van het milieu en/of verondieping in de bekkencentra aangeven en door afzettingshiaten aan hun randen. Deze facies en hiaten geven belangrijke tijdelijke veranderingen in het lokale, mediterrane of globale milieu aan. Tektonische bewegingen gerelateerd aan horizontale verschuivings bewegingen kwamen regelmatig voor in de OAP gedurende het Neogeen-Kwartair en spelen eerst een rol. Daarnaast werden 'regressieve facies' en hiaten gevormd gedurende verschillende perioden van globaal verlaagd zeeniveau. Gedurende het Messinien ontstonden dergelijke verschijnselen door een combinatie van geringe waterdiepte boven het GBDS, als gevolg van beide laatstgenoemde factoren, en sterke verdamping gedurende perioden van droog Mediterraan klimaat.

Gedurende enkele van de regressieve episoden uit het Laat Cenozoicum was de ontwikkeling van 'randbekken' en continentale omstandigheden bijzonder markant. De jongste, midden to laat Kwartair, wordt gekenmerkt door overheersende erosie in zuidoost Spanje, in de eerste plaats als gevolg van recente opheffing van de Gibraltar Boog die tot ongeveer 1000 meter oploopt in centrale delen van het orogeen. Dit verschijnsel is, in combinatie met relatief droog mediterraaan klimaat, verantwoordelijk voor de tegenwoordige geringe isolatie van de Middellandse Zee met zijn karakteristiek enigszins verhoogd zoutgehalte, betrekkelijk laag nutrient en hoog zuurstof gehalte en 'anti-estuariene' (ondiep in/diep uit) wateruitwisseling met de Atlantische Oceaan. Een tweede belangrijke regressieve episode situeert zich rond de Mio-Plioceen grens en omvat het Messinien interval, wat rijk is aan afzettingen met gespecialiseerde floras en faunas, evaporieten en andere authigene mineralen en continentale afzettingen, karst verschijnselen en erosie. Een derde dergelijke episode bevindt zich waarschijnlijk aan het eind van het Midden Mioceen, wanneeer continentale en beperkt mariene afzettingen, inclusief enige evaporieten, een wijde verbreiding vertonen in het interne deel van de Betische Cordilleren.

Gedurende het grootste deel van het Mioceen waren subtropische, tot minimaal ongeveer 1000 meter diepe, open mariene omstandigheden aanwezig in ongeveer Oost-West gerichte onderling verbonden zeestraten in het Betische gebied, waarin hemipelagische mergels met frequente turbidiet inschakelingen werden afgezet. Periodiek opwellen van voedselrijk dieper water leidde tot ingeschakelde kiezelige 'hoge vruchtbaarheids afzettingen' (in het bijzonder radiolariën), terwijl in het bijzonder in de omgeving van de tegenwoordige plaats Carboneras talrijke vulkanen actief waren. Een constellatie van 'Rif-Betische Straten' onderhield de diep-oceanische omstandigheden die worden aangegeven door het regelmatig voorkomen van bepaalde benthische microorganismen, in onze bekkens en in meer centrale delen van het Mediterrrane Bekken ('Onderst Oceanisch stadium').

Het Mio/Plioceen grens interval is goed vertegenwoordigd in de OAP en is in detail bestudeerd. Het vertoont een belangrijke ontwikkeling aan randbekken omstandigheden, in het bijzonder vertegenwoordigd door een verscheidenheid aan 'lagunaire' milieus. Na nog relatief diepe en open mariene omstandigheden aan het eind van het Tortonien, is een geleidelijke enigszins fluctuerende afname van mariene invloed waargenomen, totdat het gebied volledig geisoleerd ('continentaal') wordt op het allerlaatst van het Mioceen. Hierna vond een betrekkelijk snelle terugkeer naar open, maar ondiep mariene omstandigheden plaats in het vroege Plioceeen.

Overheersend relatief diepe en open mariene omstandigheden zijn nog waargenomen in het vroegste deel van het Messinien van zuidoost Spanje ('Open Zee stadium'). De fauna uit dit interval, die zekere verschillen vertoont met de gelijktijdige NW Atlantische, en meer expliciete facies veranderingen in de centrale en oostelijke delen van het Mediterrane Bekken, geven aan dat de water uitwisseling via het MIG al minder effectief was geworden, maar ze geleek waarschijnlijk gewoonlijk nog op het moderne type (instroom van oppervlakte water en uitstroom van bodem water). Relatief lager zuurstof en hoger nutrient gehalte werden waarschijnlijk gegenereerd door periodiek hoge verdamping en sterk 'inwellen' van Atlantisch intermediair water. De primaire factor in deze ontwikkeling was een belangrijke tektonische restructurering in het laat Tortonien, die een aantal van de Gibraltar Boog zeestraten sloot, waarbij alleen een paar ondiepere 'corridors' open bleven.

In het midden van het vroeg Messinien, begonnen de bodemwaters van de bekkens van zuidoost Spanje consistent zuurstofarm te worden, terwijl tegelijkertijd verondieping en enige stijging in zoutgehalte optrad. Het late vroeg Messinien hoog produktieve 'Open Lagune stadium' wordt gekenmerkt door regelmatige 'bloei' van slechts enkele fauna en flora elementen (buliminaceeën en epiphytische benthonische foraminiferen, het koraal *Porites*, de groene alg *Halimeda*,, en enkele planktonische foraminiferen, bryozoën, mollusken, zeeëgel, diatomeën, kiezelspons en vissoorten), die een beperkt marien milieu aangeven. Relatief geringe, cyclische zeeniveau fluctuaties kunnen worden herkend in parasequenties van de bekkenranden en in fluctuaties in zout en zuurstofgehalte weerspiegeld in de microfaunas uit de bekkencentra. Vergelijkbare facies, vaak met een armere benthische fauna, zijn wijd verbreid in de top van het vroege Messinien van de meer centrale delen van het Mediterrane Bekken, bij voorbeeld in de klassieke 'Tripoli Formatie' van Sicilie.

De regressieve trend culmineerde in de afzetting van sulfaat zouten, hetgeen aangeeft dat de uitstroom over het GBDS ineffectief werd. Gedurende dit midden Messinien, 'Mariene Evaporiet stadium' overheersten matig ondiepe, hypersaliene 'half-gesloten' tot 'gesloten lagune' omstandigheden in ons gebied, maar een relatief korte epsiode met meer open mariene maar zuurstofarme condities trad nog op. De fauna en flora die het bekken gedurende dit interval bevolkten benaderen die van het voorgaande open lagune stadium, maar vertonen toch een sterkere graad van restrictie van het milieu. Afzettingen die getuigen van belangrijke bacteriële activiteit (stromatolieten, verkalkt gips, zware metalen concentraties, ?oolieten, dolomiet) bereikten hun maximum gedurende dit interval. Het grootste deel van de dikke Messinien steenzout serie van het centrale deel van het Mediterrane Bekken werd waarschijnlijk afgezet tegen het eind van deze episode.

Het volgende laat Messinien 'Meer stadium' wordt primair gekenmerkt door terrigeen-klastische depositie, karst processen en erosie. Enkele hypersaliene perioden zijn nog waargenomen, maar substantiële evaporiet afzetting vond niet langer plaats in zuidoost Spanje en het water was, periodiek, hyperalkalien. Een paar extreem zwakke mariene 'influxen' vonden waarschijnlijk nog plaats, in het bijzonder in de basis van dit interval, maar de lage effectiviteit van de mariene verbinding wordt aangegeven door de virtuele afwezigheid van typisch marine fauna en flora. Continentale omstandigheden zijn het meest duidelijk in regelmatige rivier inschakelingen, caliche bodem vorming, karst verschijnselen en overvloedige aanwezigheid van *Microcodium*. Aanwijzingen voor matig hypersaliene omstandigheden nemen af naar de top van dit interval, waar brakke tot oligohaliene biota (*Chara*, ligniet, verschillende mollusken en ostracoden taxa) in belang toenemen, terwijl het zuurstofgehalte op de bodem verbetert. Klimaatscontrasten waren relatief sterk, waarbij de calichebodem horizons warme en droge omstandigheden weerspiegelen, terwijl wijd verbreide meer-afzettingen en karstverschijnselen relatief vochtige omstandigheden suggereren.

Dit interval correleert met de zogenaamde 'lago mare stadium' van het Mediterrane Bekken, wat vergelijkbare facies aspekten vertoont. Periodieke indamping en waarschijnlijk één of meerdere episodes van complete droogval van de Middelllandse Zee vonden plaats gedurende dit interval. De isolatie van het Mediterrane Bekken gedurende deze epsiode wordt beschouwd als het resultaat van tektonische en sedimentaire processen in het Ingangs Gebied en laag niveau van de wereldzee gedurende een ijstijd in het allerjongste Mioceen.

Aan het begin van het Plioceen, worden de diepste bekkendelen van zuidoost Spanje plotseling weer bedekt met marien water. Dit wordt primair beschouwd als het resultaat van tektonische processen in de Boog van Gibraltar, die leidden tot het 'opengaan' van de Straat van Gibraltar. Mariene omstandigheden waren aanvankelijk nog niet geheel normaal. In het bijzonder het zuurstofgehalte op de zeebodem was nog betrekkelijk laag. Vergelijkbare omstandigheden worden gevonden in de meer centrale delen van de Midddelllandse Zee hetgeen nog gedeeltelijke isolatie van het bekken suggereert. Pas enkele honderdduizenden jaren later keerden volledig open mariene omstandigheden terug in zuidoost Spanje en de centrale delen van het Mediterrane Bekken waren weer 'diep oceanisch', hetgeen het bestaan van een diepe Gibraltar doorgang impliceert ('Bovenste Oceaan stadium'). Het meest opvallende verschil met pre-Messinien omstandigheden is het verdwijnen van bepaalde subtropische fauna-elementen zoals de planktonische foraminifeer *Globorotalia menardii* en rifbouwende koralen. Terwijl de Mio-Plioceen grens wat dit aspect betreft een 'uitstervings event' vertegenwoordigt in het Mediterrane gebied, lijkt ze elders vooral gekenmerkt te worden door speciatie (verschijnen van een aantal moderne planktonische foraminiferen soorten en grotere zoogdier families als de Elephantidae en Hominidae).

Terwijl bepaalde tegenwoordig diepe delen van het Mediterrane Bekken, zoals de Alboran Zee, werden gekenmerkt door sterke daling, kwam ons gebied boven water als gevolg van zwakke daling, hoge afzettingssnelheden en opheffing gedurende het grootste deel van het Plioceeen en Kwartair.

Samenvattend geeft de facies ontwikkeling in Spanje het volgende aan voor het Mediterrane Bekken: a) verdwijnen van het 'oceaan milieu' maar nog een redelijk goede mariene verbinding gedurende het laatste Tortoon-vroegste Messinien; b) verdere beperking van de water-uitwisseling en degradatie van het diep-benthische milieu, maar grote bloei van mariene organismen in de hogere water lagen, als gevolg van een combinatie van beginnende dichtheids-stratificatie en hoge nutrient aanvoer; c) een sterke stijging van het zoutgehalte leidend tot afzetting van sulfaten aan de bekkenranden, gecombineerd met massieve zoutafzetting in diep water in het bekkencentrum in een laat stadium, nog onder in essentie continu mariene instroom condities; d) complete blokkade van de mariene instroom vanuit de Atlantische Oceaan en terrigene en evaporietafzetting in een middelmatig diepe en periodiek droogvallende 'lago mare'; e) het opengaan van een goede mariene verbinding in het vroegste Plioceen; f) tektonische modificaties in de loop van het Plioceen en in het bijzonder het Kwartair.

De aanzet voor Mediterrane isolatie werd gegeven door een belangrijke paleogeografische restructurering van haar Ingangs Gebied gedurende het late Tortoon, een gevolg van geodynamische veranderingen die een eind maakte aan de actieve daling van de Miocene bekkens. Na deze gebeurtenis geraakte de OAP in zijn 'verlandings stadium', hetgeen bestond in een vootdurende tendentie tot opheffing van de bekkenranden en opvulling van de bekken centra.

RESUME

La stratigraphie, la micropaléontologie et la géochimie des sédiments néogènes de la Province d'Alméria Orientale ('PAO', SE de l'Espagne) fournissent des données nouvelles sur l'évolution des environnements local, régional (méditerranéen) et mondial à partir d'une localité qui est bien située pour l'étude de leur interaction. La présente étude est centrée sur l'intervalle messinien qui marque la limite entre les époques du Miocène et du Pliocène, et qui est caractérisée par l'Evenement dit 'La Crise de Salinité' du Bassin Méditerranéen, où divers conditions marins 'marginaux' et continentales se développent. Les sédiments observées dans les parties plus centrales du Bassin de Nijar Septentrional (ou de Carboneras) et quelques coupes d'intervalles critiques dans les bassins de Agua Amarga, Sorbas et Vera, adjacents, ont étés étudiés en détail. Dans le but d'établir des comparaisons, des travaux de réconnaissance ont été réalisés dans les es dépôts des bordures des bassins du PAO et dans quelques coupes des parties plus externes de l'Arc de Gibraltar et en Méditerranée centrale.

La Province d'Almeria Orientale est située sur la bordure ouest du Bassin Méditerranéen, à proximité relative du Détroit de Gibraltar qui présente actuellement la seule connection de la Mer méditerranéenne avec l'Océan ouverte. Pendant le Miocène, plusieurs connections existaient au sein de cette région dite des 'Entrées Méditerranéennes' (REM), qui se trouve marquée par la présence de l'orogène alpin arqué des Bétiques-Rif-Kabylie ou Arc de Gibraltar sud-ibérique et nord-maghrebienne. La constitution (paléo)géographique de ce 'Sytème de Seuil de l'Arc de Gibraltar' (SSAG) est un facteur important dans le développement des échanges de l'eau entre la Méditerranée et l'Atlantique.

Au sein du Néogène de la Province d'Alméria Orientale un certain nombre d'unités de facies ont été reconnues sur la base de changements verticaux systématiques dans le contenu en microfossiles et les caractéristiques primaires et diagénétiques des lithologies au centre des bassins, et des changements dans le caractère lithologique, souvent associés à des surfaces d'érosion sur les bordures. Ces unités de facies sont interpretées comme représentatives d'un nombre d''épisodes du milieu' successives du région, tandis que leurs limites sont considérées comme 'évenements'. Les épisodes et les évenements ont été datées sur la base de la biostratigraphie des foraminifères planctoniques. L'évolution de l'environnement sur une période plus longue est appréciée par la comparaison des unités de facies représentant les dépôts de l'intervalle du haut-niveau marin, relativement stable, des séquences de dépôt successives.

Les dominances de la sédimentation marine profonde pendant le Miocène, de la sédimentation marine peu profonde pendant le Pliocène et de la sédimentation continentale pendant le Quaternaire, sont ponctuées par des 'épisodes régressifs' plus courts, marqués par des facies indiquant le confinement et/ou la diminution de profondeur dans le centre des bassins et par des surfaces d'érosion sur leurs bordures. Ces facies et discontinuités indiquent des changements temporaires importants dans l'environnement local, méditerranéen ou mondial. Des mouvements tectoniques en relation avec des failles transformantes, ont été fréquents dans le POA pendant le Néogène et le Quaternaire. Des 'facies regressifs' et des discontinuités ont par ailleurs été créés par divers mouvements eustatiques négatifs notables. Pendant le Messinien, de tels phénomènes se sont produits par combinaison d'une profondeur limitée à la hauteur du seuil du Bassin Méditerranéen, en relation avec ces deux derniers facteurs, avec une forte évaporation sous climat régional sec.

Quelques épisodes régressifs du Cénozoïque récent ont été particulièrement important dans la création des environnements de dépôt des 'bassins marginaux' et continentaux. Le plus récent, du Quaternaire moyen à supérieur, est caractérisé par une érosion prédominante comme résultat majeur du soulèvement récent de la région de l'Arc de Gibraltar, qui peut atteindre la valeur d'environ 1000 mètres dans des parties axiales. Ce phénomène, combiné avec un climat régional relativement sec, est responsable du léger confinement actuel du

Bassin Méditerranéen, avec ses caractéristiques particulières: salinité accrue, contenu en élements nutritifs relativement bas, oxygénation relativement élevée, eaux profondes chaudes, et circulation de l'échange d'eau 'anti-estuarienne' (entrée peu profonde/sortie profonde) avec l'Océan Atlantique. Un deuxième épisode régressif important est situé à la limite Mio-Pliocène et intéresse l'intervalle Messinien. Cet intervalle est riche en sédiments marins avec des faunes et flores spécialisées, des évaporites et autres minéraux authigènes, et des sédiments continentaux, des phénomènes de karst et d'érosion. Un troisième épisode s'est probablement produit vers la fin du Miocène moyen, quand la sédimentation continentale ou marine confinée, incluant quelques évaporites, s'est répandue dans la partie interne des Cordillères Bétiques.

Pendant la majeure partie du Miocène, des conditions de mer ouverte, subtropicale-chaude, avec des profondeurs atteignant au moins ca. 1000 mètres, étaient réalisées dans la région bétique au sein de sillons allongés selon une direction approximativement E-W, dans lequels se sont déposées des marnes hémipélagiques avec de fréquents intercalations turbiditiques. L'introduction périodique d'eaux riches en éléments nutritifs par des courants de 'upwelling', a conduit à des dépôts de haute-fertilité (riches en radiolaridés), tandis que, en particulier dans les environs de Carboneras, de nombreux volcans étaient actifs. Un multitude de 'Détroits betico-riffains' ont permis le maintien de conditions océaniques profondes, qui sont indiquées par des microfaunes benthiques dans nos bassins ainsi que dans les parties plus centrales du Bassin Méditerranéen ('Episode océanique inférieur').

L'intervalle de transition Mio/Pliocène est bien représenté dans le PAO et a été étudié en détail. Il montre un développement important des conditions de 'bassin marginal', marquées en particulier par une diversification des environnements 'lagunaires'. Au milieu encore relativement profond et ouvert pendant le Tortonien terminal, fait suite une période soulignée, au sommet du Miocène, par une diminution graduelle, quelque peu fluctuante, des influences marines conduisant à l'isolement complète de la région, ainsi 'continentalisée'. Par la suite, un retour relativement rapide à des conditions de mer ouverte, mais moins profonde, s'est produit pendant le début du Pliocène.

Des conditions dominantes de mer relativement profonde et ouverte sont encore enregistrées dans la partie la plus ancienne du Messinien du Sud Est de l'Espagne ('Episode de mer ouverte'). La faune de cet intervalle, qui montre certaines différences avec la NW atlantique contemporaine et des changements de facies mieux exprimés dans la partie centrale et orientale du Bassin Méditerranéen, indiquent que l'échange des eaux a travers le REM était déja devenu moins effectif. Cette situation évoquait probablement le type actuel avec entrée superficielle et sortie des eaux en profondeu durant la plus grande partie de cet épisode. Un milieu quelque peu réduit en oxygène et relativement riche en éléments nutritifs étaient probablement génerés par une haute évaporation périodique avec 'inwelling' des eaux Atlantiques intermédiaires. Le facteur primaire dans ce développement était une importante restructuration tectonique dans le Tortonien terminal, qui conduisait à la fermeture de plusieurs des détroits de l'Arc de Gibraltar. Seuls quelques corridors peu profonds subsistèrent.

Au milieu du Messinien inférieur, les eaux profondes des bassins de l'Espagne du Sud Est ont commencé à présenter un déficit en oxygène continu, s'accompagnant d'une diminution de profondeur et d'une augmentation modérée de la salinité. 'L'Episode de lagon ouvert', fortement productif, du Messinien inférieur terminal est caractérisé par la haute fertilité des élements fauniques at floristiques relativement peu diversifiés (des foraminifères benthiques buliminacés et epiphytiques, le corail *Porites* , l'algue verte *Halimeda*, et quelques espèces de foraminifères planctoniques, bryozoaires, diatomées, éponges silicieuses et poissons), qui indiquent un environnement marin confiné. Des fluctations cycliques mineures du niveau de la mer peuvent être reconnues dans des paraséquences développées sur les bordures des bassins et dans des variations de la salinité et du contenu en oxygène réflètées par les microfaunes du bassin. Des facies comparables, mais plus pauvres en biota benthiques, sont répandus dans le Messinien inférieur terminal de la Méditerranée Centrale, par exemple dans la Formation Tripoli, classique, de la Sicile.

La tendance régressive culmina avec le dépôt des évaporites sulfatées, ce qui indique que le rejet des eaux profondes du bassin à travers le SBAG était au moins périodiquement interrompu. Pendant cet 'Episode des Evaporites Marines' du Messinien moyen, des conditions hypersalines, assez peu profondes, d'un 'lagon sémi-restreint' à 'restreint' étaient dominantes dans notre région. Un épisode relativement court induisant des conditions marines plus ouvertes, mais à déficit en oxygène, s'est toutefois encore produit. Les faunes qui habitaient le bassin pendant cet intervalle sont proches de celles de l'épisode de lagon ouvert précédant, mais montrent cependant un degré de confinement plus fort. Des dépôts qui témoignent d'une activité microbienne importante (stromatolites, gypses calcifiés, concentrations de métaux lourds, ? oolites, dolomites), atteignent leur maximum. La majeure partie des séries épaisses du sel messinien de la Méditerranée Centrale s'est probablement déposée à la fin de cet épisode.

'L'episode lacustre' du Messinien supérieur qui suit est caractérisé tout d'abord par une déposition clastique-terrigène, une karstification et de l'érosion. Quelques sédiments hypersalins sont encore enrégistrés, mais le dépôt substantiel des évaporites se ne produit plus dans le SE de l'Espagne et, périodiquement, les eaux deviennent hyperalcalines. Quelques 'entrées' marines extrêmement faibles se sont probablement encore produites, en particulier vers la base de cet intervalle, mais la faible probabilité de connection marine est indiquée par l'absence virtuelle de flore et de faune marines. Les environnements continentaux les plus évidents sont matérialisés par des intercalations fluviatiles fréquentes, la formation de sols calcrétiques, les phénomènes de karstification et l'abondance de *Microcodium*. Les indications en faveur de conditions moyennement hypersalines décroisssent vers le sommet de cet intervalle, où des biota saumâtres à oligohalins (*Chara*, lignite, divers mollusques et ostracodes) deviennent plus importants, tandis que l'oxygénisation s'améliore. Des contrastes climatiques forts ont dû exister, les horizons des sols calcrétiques attestant des conditions chaudes et sèches, tandis que les facies lacustres et la karstification repandus suggèrent une humidité relative.

Cet épisode est corrélable avec la phase de 'lac mer' de la Méditerranée Centrale qui présente des facies similaires. Des abaissements évaporitiques périodiques et, probablement, des dessications complètes de la 'Mer' méditerranéenne se sont produits pendant cet intervalle. L'isolation du Bassin Méditerranéen pendant cet épisode est considérée comme résultant des processus tectoniques et sédimentaires dans la REM et d'un bas niveau marin mondial pendant un événement glaciaire au sommet même du Miocène.

Au début du Pliocène, les bassins plus profonds du SE de l'Espagne sont soudainement inondés par des eaux marines. Ceci est tout d'abord le résultat des processus tectoniques dans l'Arc de Gibraltar, qui ont entrainé 'l'ouverture' du Détroit de Gibraltar. Initialement, pendant le Pliocène inférieur basal, les conditions marines n'étaient pas encore complètement normales. Le contenu en oxygène sur les fond de la mer en particulier était encore relativement faible. Des phénomènes comparables sont rencontrés dans des parties plus centrales de la Méditerranée indiquant une isolation encore partielle du bassin.

Quelques centaines de milliers d'ans après, des conditions de mer complètement ouverte étaient de nouveau établies dans le SE de l'Espagne et les parties centrales de la Méditerranée devenaient de nouveau 'océaniques', ce qui implique l'existence d'un passage profond au droit de Gibraltar ('Episode océanique supérieur'). La différence la plus évidente avec les conditions pré-messiniennes est la disparation des élements de faune marine subtropicale chaude, tels le foraminifère planctonique *G. menardii* et les coraux hermatypiques. Tandis que la limite Miocène-Pliocène est alors réprésentative d'un 'événement d'extinction' dans le Bassin Méditerranéen, elle semble ailleurs être caractérisée essentiellement par la spéciation (apparition d'un nombre des espèces de foraminifères planctoniques modernes et des familles de grands mammifères comme les Eléphantidae et les Hominidae).

Tandis que certaines parties actuellement profondes de la Méditerranée, comme la Mer d'Alboran, connaissaient une forte subsidence, notre région d'étude subissait encore une perte de profondeur et finalement une érosion, déterminées par une faible subsidence, des taux de sédimentation relativement élevés et par un soulèvement pendant la majeure partie du Pliocène et le Quaternaire.

En résumé, le développement des facies dans le SE de l'Espagne indique pour la Méditerranée: a) la disparition de 'l'environnement océanique', avec toutefois persistance d'une connection marine relativement effective pendant le Tortonien terminal-Messinien inférieur basal; b) une restriction accrue des échanges d'eaux marines et une dégradation de l'environnement benthique profond, mais une grande fertilité des organismes marins dans les couches d'eaux supérieures, dues à une combinaison d'initiation de la stratification de la densité et d'un fort apport en éléments nutritifs; c) un fort accroisement de la salinité sous flux marin continu. Les sulphates se déposent sur les bordures du bassin, ultérieurement s'y associent en position plus centrale sels d'eaux profondes; d) un blockage complèt de l'entrée de l'eau Atlantique avec dépôt des sédiments terrigènes et évaporitiques dans un lac-mer moyennement profond et periodiquement asséché; e) l'instauration d'une communication marine très nette dans le cours du Pliocène inférieur basal; f) ajustements tectoniques au cours du Pliocène et surtout pendant le Quaternaire.

L'isolation méditerranéenne était préparé par des réorganisations paléogéographiques importantes de sa 'Région d'Entrées' pendant le Tortonien terminal, comme résultat des changements géodynamiques, qui mirent fin à la subsidence active des bassins miocènes. Après ces événements, la PAO est entrée dans son 'épisode d'émergence', qui consista en une tendence continue vers le soulèvement des marges des bassins et au comblement de leurs parties centrales.

SUMMARY

The stratigraphy, micropaleontology and geochemistry of Neogene sediments of the Eastern Almeria Province ('EAP', SE Spain) provide new data on the evolution of the local, regional (Mediterranean) and global environment from an area which is well-suited to study their interaction. The present study emphazises the Messinian interval, marking the boundary between Miocene and Pliocene Epochs, and characterized by the so-called 'Salinity Crisis' in the Mediterranean Basin, where a variety of 'marginal marine' and continental conditions temporarily developed. The sediments in the more central part of the Northern Nijar or Carboneras Basin and some additional sections of critical intervals in the adjacent Agua Amarga, Sorbas and Vera basins have been studied in detail. For comparison, reconnaisance studies were made of the deposits of the basin margins of the EAP, and of some sections in more external parts of the Gibraltar Arc and in the central Mediterranean.

The Eastern Almeria Province is located at the western margin of the Mediterranean Basin, relatively close to the Strait of Gibraltar, which presently is the only connection of the Mediterranean Sea with the open ocean. During the Miocene several connections existed in the 'Mediterranean Entrance Area' (MEA), which is marked by the presence of the arcuate, alpine Betic-Rif-Kabylian orogen or Gibraltar Arc of southern Spain and the northernmost Maghreb. The (paleo)geographic constitution of this 'Gibraltar Arc Sill System' (GASS) is an important factor in the effectivity of the water-exchange between Mediterranean and Atlantic.

Within the Neogene of the Eastern Almeria Province a number of subsequent facies units have been recognized on basis of consistent vertical changes in microfossil content and primary and diagenetic lithologic characteristics in the basin centres, and changes in lithologic character, often marked by erosional surfaces, at their margins. These facies units are interpreted in terms of successive 'environmental stages' of the area, whereas major facies boundaries are numbered as 'events'. Stages and events have been dated primarily by means of planktonic foraminifer biostratigraphy. The environmental development over a longer time interval is best studied by comparing facies units representing the relatively stable highstand deposits of successive depositional sequences.

The subsequent dominance of deep marine sedimentation during the Miocene, of shallow marine sedimentation during the Pliocene and of continental sedimentation during the Quaternary, is punctuated by shorter 'regressive episodes', marked by facies indicating restriction and/or shallowing in the basin centres and unconformities at their margins. These facies and unconformities indicate important temporary changes in the local , mediterranean or global environment. Tectonic movements related to strike-slip faults have been common in the EAP during the Neogene-Quaternary and first play a role. Besides, 'regressive facies' and unconformities were created during several important global sealevel lowerings. During the Messinian, such features originated by a combination of shallow waterdepths over the GASS, due to both latter factors, and strong evaporation during dry regional climate.

A few regressive episodes of the Late Cenozoic were particularly important in creating 'marginal basin' and continental conditions. The youngest, middle to late Quaternary, is characterized by predominant erosion in SE Spain, primarily due to young uplift of the Gibraltar Arc, which amounts to ca. 1000 metres in axial parts. This phenomenon, in combination with relatively dry regional climate, is responsable for the present slight isolation of the Mediterranean Basin with its particular characteristics of somewhat increased salinity, relatively low nutrient and high oxygen content, warm bottom waters and 'anti-estuarine' (shallow in/deep out) water-exchange with the Atlantic Ocean. A second, important regressive episode is located around the Mio-Pliocene boundary and comprises the Messinian interval, which is rich in marine deposits with specialized floras and faunas,

evaporites and other authigenic minerals and continental deposits, karst and erosion. A third such episode probably occurred at the end of Middle Miocene, when continental and restricted marine deposits, including some evaporites, are widespread in the internal part of the Betic Cordilleras.

During the main part of the Miocene, warm-subtropical, up to at least ca. 1000 m deep, open marine conditions existed in approximately E-W trending interconnected straits in the Betic area in which hemipelagic marls with common turbidite intercalations were deposited. Periodic introduction of nutrient-rich waters by upwelling, lead to intermittent siliceous high-fertility deposits (especially rich in radiolarians), whereas in particular in the vicinity of Carboneras, numerous volcanoes were active. A constellation of 'Rif-Betic Straits' sustained the deep, oceanic conditions that are recorded by smaller benthic organisms, in our basins and more central parts of the Mediterranean ('Lower Oceanic stage').

The Mio/Pliocene transition interval is well represented in the EAP and has been studied in detail. It shows an important development of marginal basin conditions, in particular represented by a variety of 'lagoonal' environments After still relatively deep, open marine conditions in the latest Tortonian, a gradual, somewhat fluctuating decrease of marine influence is recorded, untill the area becomes completely isolated ('continental') at the very top of the Miocene. Hereafter, a relatively rapid return to open, but shallow marine conditions occurrred in the early Pliocene.

Dominant, relatively deep and open marine conditions are still recorded in the earliest part of the Messinian of SE Spain ('Open sea stage'). The faunas from this interval, which show certain differences with the contemporaneous NW Atlantic ones, and more explicit facies changes in the central and eastern parts of the Mediterranean Basin indicate that water-exchange through the MEA had already become less effective but it probably still commonly resembled the modern type (surficial inflow and outflow of bottom waters). Relatively lower oxygen and higher nutrient content were probably generated by periodic high evaporation and strong 'inwelling' of Atlantic intermediate water. The primary factor in this development was an important tectonic restructuring in the late Tortonian, which closed a number of the former Gibraltar Arc seaways with only a few shallower corridors remaining.

In the middle of the early Messinian, the bottom waters of the basins in SE Spain began to become consistently oxygen-deficient, which was accompanied by shoaling and some rise in salinity. The late early Messinian, highly productive 'Open lagoon stage' is characterized by common high-fertility of relatively few faunal and floral elements (buliminacean and epiphytic benthic foraminifers, the coral *Porites*, the green algae *Halimeda*, and a few planktonic foraminifer, bryozoan, diatom, siliceous sponge and fish species), indicating a restricted marine environment. Relatively minor, cyclic sealevel fluctuations can be recognized in parasequences developed at the basin margins and in variations in salinity and oxygen content reflected in the basinal microfaunas. Comparable facies, with an often poorer benthic biota content, are widespread in the late early Messinian of the Central Mediterranean, for instance in the classical Tripoli Formation of Sicily.

The regressive trend, culminated in the deposition of sulphate evaporites, indicating that outflow over the GASS became ineffective. During this middle Messinian, 'Marine Evaporite stage', moderately shallow, hypersaline 'semi-restricted' to 'restricted lagoon' conditions were dominant in our area, but a relatively short, more open marine, but oxygen deficient episode still occurred. The faunas and floras that populated the basin during this interval approach those of the preceeding open lagoon stage but show a still stronger degree of confinement. Deposits attesting to important microbial activity (stromatolites, calcitized gypsum, base metal concentration levels, ?oolites, dolomite) reached their maximum during this interval. The main part of the thick Messinian salt deposits of the Central Mediterranean were probably deposited at the end of this episode.

The following late Messinian "Lacustrine stage" is primarily characterized by clastic deposition, karstification and erosion. A few hypersaline periods are still recorded but substantial evaporite deposition no longer took place in SE Spain and, periodically, the water was hyperalkaline. A few extremely weak marine 'influxes' probably still occurred, in particular near the base of this interval, but the low effectivity of the marine connection is indicated by the virtual absence of typical marine flora and fauna. Continental conditions are most obvious in common river intercalations, caliche soil formation, karst features and the abundance of *Microcodium*. Indications of moderately hypersaline conditions decrease towards the top of this interval, where brackish to oligohaline biota (*Chara*, lignite, several mollusk and ostracode taxa) increase in importance, whereas oxygenation improved. Climatic contrasts were relatively strong, the calcrete horizons attesting to warm and dry conditions, whereas the widespread lacustrine facies and karstification suggest relative humidity.

This episode correlates with the so-called 'lago mare stage' of the Central Mediterranean which shows similar facies aspects. Periodic evaporitic drawdown, and, probably, complete dessication of the Mediterranean 'Sea' occurred during this interval. The isolation of the Mediterranean Basin during this episode is considered the result of tectonic and sedimentary processes in the MEA and low global sealevel during a glaciation event at the very end of the Miocene.

At the start of the Pliocene, the deepest basinal areas of SE Spain are suddenly reflooded by marine waters. This is primarily the result of tectonic processes in the Gibraltar Arc, which led to the 'opening' of the Strait of Gibraltar. Initially, marine conditions still were not completely normal. In particular oxygen content at the seabottom was relatively low. Similar features are found in the Central Mediterranean, suggesting still partial isolation of the basin. Only some hundred thousand years later, fully open marine conditions returned in SE Spain and the central parts of the Mediterranean were 'deep oceanic' again, which implicates the existence of a deep Gibraltar passage ('Upper Oceanic stage'). The most obvious difference with pre-Messinian conditions is the disappearance of certain subtropical marine faunal elements such as the planktonic foram *G. menardii* and reef-building corals. Whereas the Mio-Pliocene boundary thus represents an 'extinction event' in the Mediterranean, it elsewhere seems to be primarily characterized by speciation (appearance of a number of modern planktonic foram species and larger mammals families as elephantids and hominids).

Whereas strong subsidence characterized certain presently deep parts of the Mediterranean, such as the Sea of Alboran, our study area emerged as the result of weak subsidence, relatively high sedimentation rates and uplift during the main part of the Pliocene and Quaternary.

Summarizing, the facies development in SE Spain indicates for the Mediterranean: a) disappearance of the 'ocean environment' but a still relatively good marine connection during the latest Tortonian-earliest Messinian; b) further restriction of water-exchange and degrading of the deep benthic environment, but high fertility of marine organisms in the higher water layers, due to a combination of intiating density stratification and high nutrient supply; c) a strong rise in salinity leading to marginal sulfate deposition, which was combined with massive salt deposition in deep water in the basin centre in a late stage, still under predominant conditions of continuous inflow; d) complete blocking of Atlantic inflow and terrigenous and evaporite deposition in a moderately deep and periodically dessicated 'lago mare'; e) the opening of a good connection in the early Pliocene; f) tectonic readjustments in the course of the Pliocene and especially Quaternary.

The stage for Mediterranean isolation was set by important paleogeographic rearrangements of its Entrance Area during the late Tortonian, a result of geodynamic changes, which ended active subsidence of the Miocene basins. After this event, the EAP entered into its 'emergent stage', which consisted in a continuous tendency to uplift of the basin margins and infill in the basins proper.

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CHAPTER 1

INTRODUCTION

STUDY OUTLINE AND PREVIOUS WORK

This thesis concerns Mio-Pliocene sediments of the Eastern Almeria Province ('EAP'), SE Spain. Their, diversity, high degree of exposure, relatively mild tectonization and location, make them well-suited for basin analysis by means of paleontology, geochemistry, sedimentology and stratigraphy, as well as for the study of subjects of more general geologic interest. This study concentrates on the sediments deposited under 'marginal basin' and continental conditions, which are widespread in the Messinian, a time-interval when the Mediterranean constituted a restricted part of the open-marine environment, which developed towards continental near its top.

Our data were assembled in order to test hypothesis concerning two specific geologic subjects. The first is the 'Messinian Event', which marks the boundary between the Miocene and Pliocene Epochs and is best known for the large amounts of evaporites deposited in the Mediterranean, commonly thought to have been accompanied by one or more complete dessications of this basin. Earlier studies, by Ryan (1973), Van Couvering et al (1976), Cita & Ryan (1978, 1979) and Van Hinte (1982, 1990, 1991) a.o., propose that the Messinian evolution of the Mediterranean and its Entrance Area lead to global events. Because of its location at the Mediterranean side of the present Strait of Gibraltar, and the diverse connections that existed through the Betic-Rif ('Gibraltar Arc') orogen in Miocene times (e.g. Benson 1976, Van Couvering et al. 1976, Montenat 1977, Müller & Hsü 1987, Benson et al. 1991; Figs 1.1 & 2.1), the Almeria area is (1) well-suited to study the history of the Mediterranean inlets, whereas (2) its facies and stratigraphic development record the environmental development within the Mediterranean Basin.

The second subject is the young geologic history of the Betic Cordilleras of southern Spain, where, during most of the Neogene, transcurrent movements played an important role in the late stage of the evolution of an alpine mountain range (e.g. Hermes 1985, Montenat et al. 1987, 1990, Sanz de Galdeano 1990a, De Jong 1991).

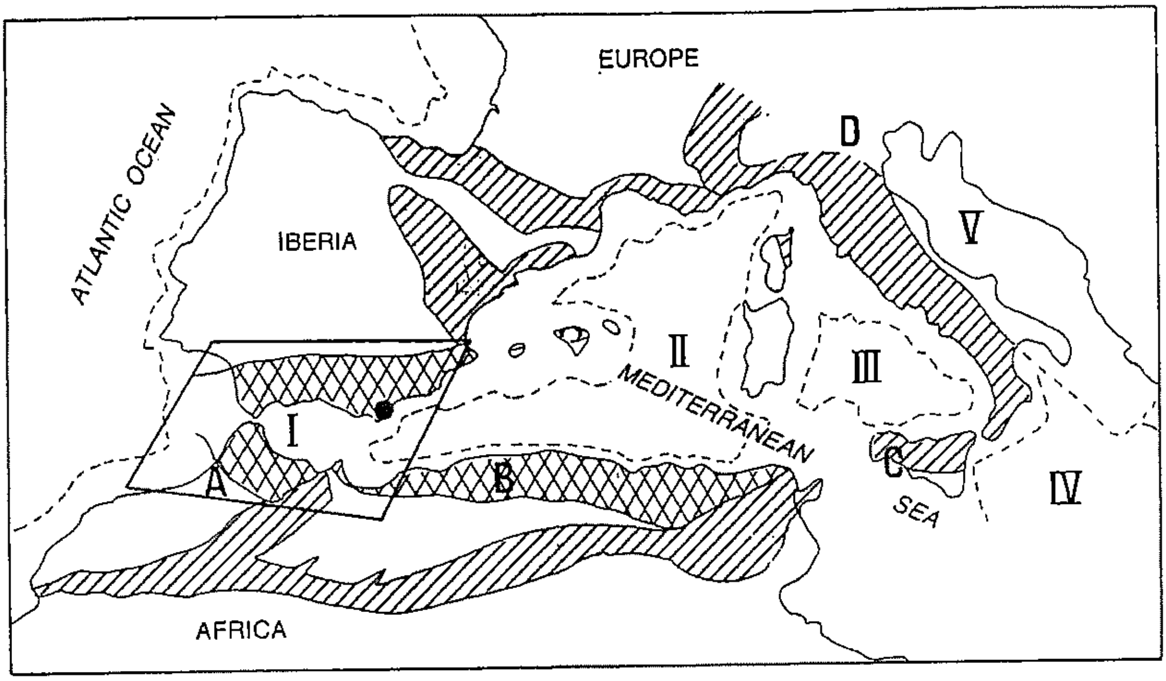


FIGURE 1.1. Map of the western and central Mediterranean Basin showing: 1) its enclosure by numerous alpine mountain chains (hatched), 2) its Entrance Area ('MEA', encadrated), with the present narrow connection with the open ocean (Strait of Gibraltar) in the Gibraltar Arc mountain system (cross-hatched) 3) the location of the main study area, the Eastern Almeria Province (EAP, black dot)), 4) its main present subbasins (Alboran-I; Balearic or Algéro-Provençal-II; Tyrrhenian-III; Ionian-IV and Adriatic-V), 5) some other presently elevated areas with well-developed Mio-Pliocene deposits discussed in the text (Gharb Basin-A; Chelif Basin-B; Central Sicilian Basin-C; Periadriatic/Po Basin-D).

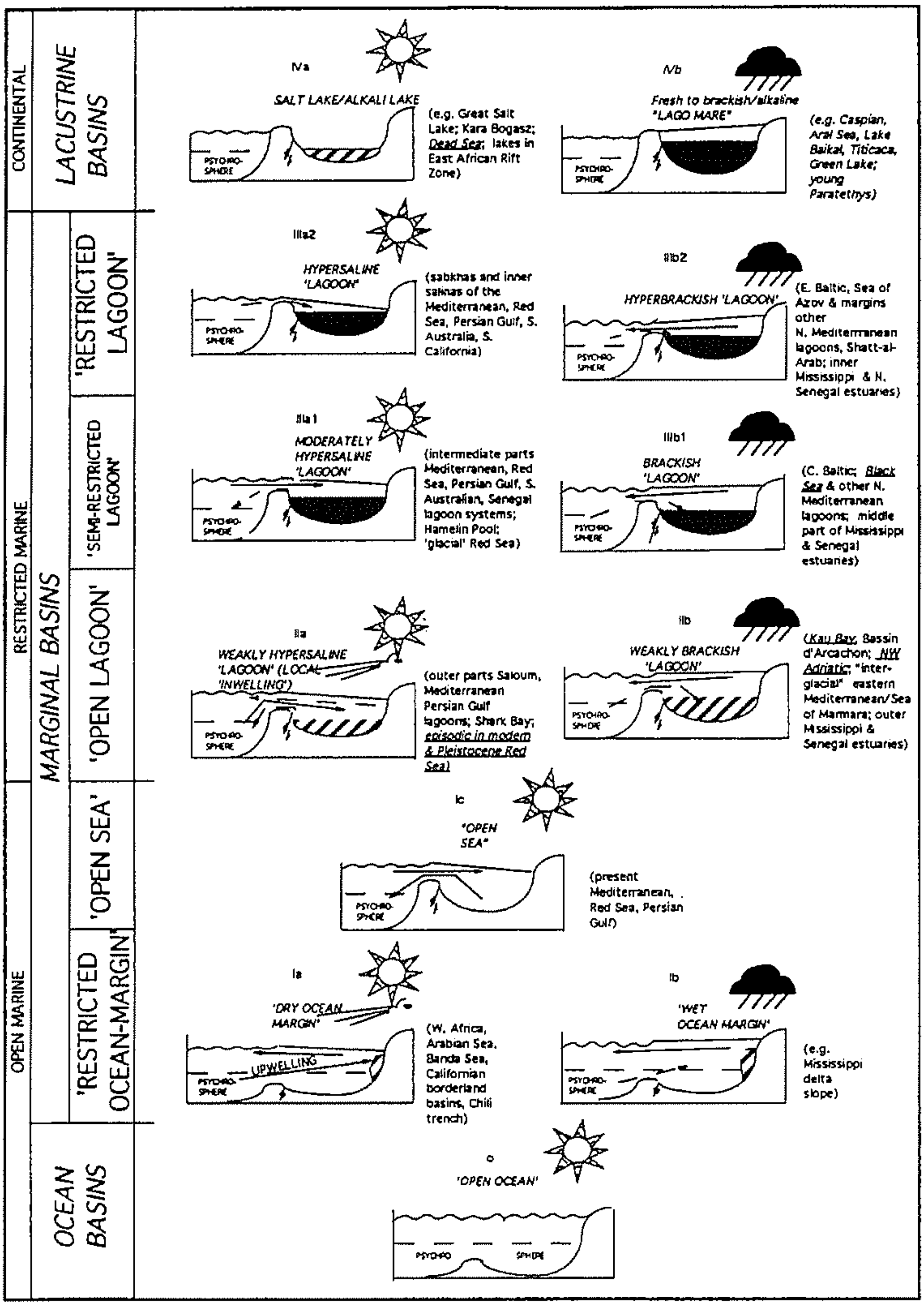
The dashed line marks the boundary between continental and oceanic crust and approximates the 2500 m isobath in the Mediterranean, which marks the upper limit of thick Messinian salt deposits as far as the Balearic and Ionian subbasins are concerned (after Cita et al. 1978 and Durand Delga & Fontboté 1980).).

MARGINAL BASIN FACIES

'Marginal basins' (more or less enclosed parts of the marine environment under strong influence of continental climate; Figs 1.1 & 1.2), often are the locus for accumulation of a variety of economically valuable minerals (hydrocarbons, sedimentary ores, diverse salt minerals), whereas common diagenetic events may produce beds with good reservoir properties (e.g. Schmalz 1969, Demaisson & Moore 1980, Friedman 1980, Arthur et al. 1987, Pierre & Rouchy 1988, Tannenbaum & Charrach 1993). The study of geologic examples of such basins has predictive value for environmental studies of their modern counterparts which are commonly situated in densely populated areas, or have interest from the viewpoint of tourist or seafood-industries (e.g. Caspers 1957, Zenkevich 1957, Harbridge et al. 1976, Carbonel et al. 1981a, Masse 1987, Barnabé 1989, Boomer 1993, Sinninghe Damsté et al. 1993).

The signal of geologic changes in global sealevel or climate is often amplified in such enclosed areas (e.g. Thunnell et al. 1987, 1991), which will be equally sensitive to man-induced environmental changes. The development of such satellite basins may also exert an important influence on the open ocean environment and global climate by the temporary accumulation, alteration and subsequent massive injection of certain of its chemical components (Ryan 1973, Ryan et al. 1974, Thierstein & Berger 1978, Vincent et al. 1980, Blanc & Duplessey 1982, Van Hinte 1982, 1990, 1992, Thunnell et al. 1987, Beekman 1992).

Models for marginal basin development with description of characteristic environmental parameters and facies have been given by amongst others Schmalz (1969), Brongersma Sanders (1970) and Dronkert (1985) with emphasis on deep, and by Perthuisot & Guelorget (1987) with emphasis on shallow basinal settings. Each type of marginal basin situation is characterized by specific values of a number of important environmental parameters (salinity, oxygen and nutrient content, temperature) (cf. Fig. 1.2). Differences in composition and abundance of biota and authigenic minerals represent variations in these parameters but the relative degree of confinement also plays an important role in their development (Perthuisot & Guelorget 1987). Relative sill-depth and regional climate are major factors in the facies development of a marginal basin, whereas the dimensions of the basin also plays an important role as especially underlined by Schmalz (1969).


CONTINENTAL
LACUSTRINE BASINS
RESTRICTED MARINE
MARGINAL BASINS
'RESTRICTED LAGOON'
'SEMI-RESTRICTED LAGOON'
'OPEN LAGOON'
OPEN MARINE
'OPEN SEA'
'RESTRICTED OCEAN-MARGIN'
OCEAN BASINS
IVa
SALT LAKE/ALKALI LAKE
(e.g. Great Salt Lake; Kara Bogasz; Dead Sea; lakes in East African Rift Zone)
IVb
Fresh to brackish/alkaline "LAGO MARE"
(e.g. Caspian, Aral Sea, Lake Baikal, Titicaca, Green Lake; young Paratethys)
IIIa2
HYPERSALINE 'LAGOON'
(sabkhas and inner salinas of the Mediterranean, Red Sea, Persian Gulf, S. Australia, S. California)
IIIb2
HYPERBRACKISH 'LAGOON'
(E. Baltic, Sea of Azov & margins other N. Mediterranean lagoons, Shatt-al-Arab; inner Mississippi & N. Senegal estuaries)
IIIa1
MODERATELY HYPERSALINE 'LAGOON'
(intermediate parts Mediterranean, Red Sea, Persian Gulf, S. Australian, Senegal lagoon systems; Hamelin Pool; 'glacial' Red Sea)
IIIb1
BRACKISH 'LAGOON'
(C. Baltic; Black Sea & other N. Mediterranean lagoons; middle part of Mississippi & Senegal estuaries)
IIa
WEAKLY HYPERSALINE 'LAGOON' (LOCAL 'INWELLING')
(outer parts Saloum, Mediterranean Persian Gulf lagoons; Shark Bay; episodic in modern & Pleistocene Red Sea)
IIb
WEAKLY BRACKISH 'LAGOON'
(Kau Bay, Bassin d'Arcachon; NW Adriatic; "inter-glacial" eastern Mediterranean/Sea of Marmara; outer Mississippi & Senegal estuaries)
Ic
"OPEN SEA"
(present Mediterranean, Red Sea, Persian Gulf)
Ia
'DRY OCEAN MARGIN'
UPWELLING
(W. Africa, Arabian Sea, Banda Sea, Californian borderland basins, Chili trench)
Ib
'WET OCEAN MARGIN'
(e.g. Mississippi delta slope)
o
'OPEN OCEAN'
PSYCHRO-SPHERE

<----- FIGURE 1.2. Cartoon illustrating deep marginal basin situations as a function of variable sill depth and regional climate. (Sub)recent modern examples (or approximates) used for comparison in this study are indicated. Hatched to blackened areas denote suboxic to anoxic bottom-water conditions. Approximate (surface layer) salinity values: 0-15 ppt for hyperbrackish, 15-25 ppt for brackish, 25-30 ppt for weakly brackish 'lagoons', 30-42 ppt for open marine conditions, and 42-50 ppt for weakly, 50-150 ppt for moderately, > 150 ppt for strongly hypersaline 'lagoons'. The term 'open lagoon' approximately corresponds to the 'lagon ouvert' of Carbonel et al. (1981).

Note importance of relative sill-depth (dependant on tectonics and, if already shallow, on sedimentary and eustatic processes) and regional climate upon quantity and quality of water and ion-exchange between continental, 'marginal basin' and open ocean environment.

Data-base: Murray 1973 (benthic foraminifers, general); Carbonnel 1969, Carbonel et al. 1981, Guernet & Lethiers 1989 (ostracodes, general); Bronstein 1947, Zenkevich 1957, Benson 1969, 1978a, Godel et al. 1971, Olteanu 1978, Carbonel & Peypouquet 1979, 1983, Peypouquet et al. 1979, 1983, Texeir et al. 1980, Boyer 1981, Mouriguiart 1987, Plaziat 1988, Olteanu & Vekua 1989, Thompson et al. 1990, Anadon 1992, Boomer 1993 (lake environments); Caspers 1957, Van Straaten 1970, Gheorgian 1974, 1978, Benson 1978b, Carbonel 1978, 1987, Gersonde 1978, 1980, Olteanu 1978, Schrader & Gersonde 1978, Sturani 1973, Liebau 1980, Sonnenfeld 1980, Stanley & Blanpied 1980, Carbonel & Pujos 1981, Marcus & Thorhaug 1981, Monteillet et al. 1982, Zaninetti & Tetart 1982, Bodergat 1983, Carbonnel 1983, Massouri et al. 1985, UNESCO 1985, Warren & Kendall 1985, Debenay et al. 1987, 1990, De Deckker et al. 1988, Plaziat 1988, Middelburg 1990, Barmawidjaja 1992, Moodley 1992 (lagoonal environments); Benson 1976, 1978a, Wright 1978a & b, Bonaduce et al. 1983a & b, Cronin 1983, Peypouquet 1983, Gersonde & Schrader 1984, Troelstra & Kroon 1989, Troelstra et al. 1989, Demarq 1990, Hasegawa et al, 1990, Benson et al. 1991, Ganssen & Kroon 1991, Abrantes 1992, Ottens 1992, Saager 1994 (open marine environments); further references in Chs. 3.2 & 3.3, Van de Poel et al. 1992, Van de Poel & Schlager, in press and Van de Poel & Boekschoten, in prep.).

THE MESSINIAN EVENT

The Messinian is, by definition, the last stage of the Miocene. Early geologic workers already suggested important environmental changes by placing the boundary between Miocene and Pliocene Epochs at its level (Lyell *in* Cita 1975a, Van Couvering et al. 1976, Van Couvering 1977). In more recent years important events have been recognized at the end of the Miocene, both within and outside the Mediterranean. The best documented event is the so-called 'Messinian Salinity Crisis of the Mediterranean' (e.g. Ruggieri 1967, Hsü et al. 1973a & b), recorded in evaporites throughout the basin, which, together with a number of associated features, are sandwiched between open marine deposits of Tortonian and Pliocene age. Cooling of ocean waters, enhanced glaciation and global (eustatic) regression, are, probably interrelated, 'Messinian events', that have been recognized outside the Mediterranean Basin (e.g. Bandy 1966, *et al.* 1969, 1972, Kennett 1967, Hayes & Frakes 1975, Shackleton & Kennett 1975, Van Couvering et al. 1976, Van Hinte 1982, Adams et al. 1977, Shackleton & Cita 1979, Van Gorsel & Troelstra 1981, Mercer & Sutter 1982, McKenzie et al. 1984, 1988, Hodell et al. 1986, 1989 a & b, Haq et al. 1987, Müller and Hsü 1987, Keigwin 1987, Benson et al.

1991). Some fundamental debate exists over the question whether it was primarily the Messinian Salinity Crisis of the Mediterranean which triggered glaciation or vice-versa (Ryan 1973, Ryan et al. 1974, Van Couvering et al. 1976, Cita & Ryan 1978, 1979, Van Hinte 1982, 1992, McKenzie et al. 1979, 1984, Müller & Hsü 1987). This situation has been referred to as 'the temperate tail wagging the polar dog' (or vice-versa) (Berggren & Haq 1976, Van Couvering et al. 1976, Cita 1977).

The idea that the whole Mediterranean Basin underwent a salinity crisis was worked out in the 1960's by Italian authors (Ruggieri 1961, 1967, Selli 1960, 1971, 1973). The determination of the Messinian age of the large amount of evaporites buried beneath the seafloor of the Mediterranean during DSDP Leg 13 (Ryan, Hsü et al. 1973), strongly supported this hypothesis, but at the same time widely divergent opinions on important details of this event were forwarded. Adherents to the idea of a salinity crisis held 3 hypotheses, namely that all the evaporites in the deeper parts of the Mediterranean were deposited in 1) a several thousands meters deep, dessicated basin, 2) a shallow dessicated basin or, 3) a basin with deep hypersaline waters (e.g. Hsü et al. 1973a and b, Nesteroff 1973, Selli 1973, Debenedetti 1976), whereas others still contended localized evaporite deposition with normal marine sedimentation continuing elsewhere in the Mediterranean (Braune et al. 1973, Meulenkamp & Zachariasse 1973, Montenat 1973b, Perconig 1973, Sturani 1973, Tauecchio & Marks 1973).

In subsequent years, the more extreme theories with respect to Mediterrranean evaporite deposition were modified into somewhat more convergent models, admitting deposition under continuous inflow and deposition in relatively deep water for the lower part of the evaporites (in particular the 'main salt unit') and admitting dessication phases during shallower deposition of its upper part (Van Couvering et al. 1976, Hsü et al. 1977, Montadert et al. 1978, Busson 1979, Dronkert et al. 1979, Rouchy & Orszag Sperber 1980; Fig. 1.3). Rouchy (1981, 1982) and Van der Zwaan (1982) further combined the most conflicting aspects of the different theories in a model with younger evaporite deposition in a basin of intermediate depth, admitting strong drawdown of the Mediterranean waterlevel and mixing of considerable amounts of fresh water with marine influxes in the upper part of the Messinian. More recent publications mainly apport more arguments for the models of the late seventies and further attempt to correlate Messinian events within the Mediterranean with global events (McKenzie et al. 1984, Müller & Hsü 1987, Busson 1990, Saint-Martin & Rouchy 1990, Benson et al. 1991), although the dessication theory has been recently again challenged by Dietz and Woodhouse (1988) who contend that the Messinian salts are not evaporites but 'precipitates'.

Most recently proposed models (Fig. 1.3; Busson 1990) agree on 1) the fact that the Mediterranean Basin became more or less isolated in latest Messinian times and of the existence of its salinity crisis 2) a two-fold

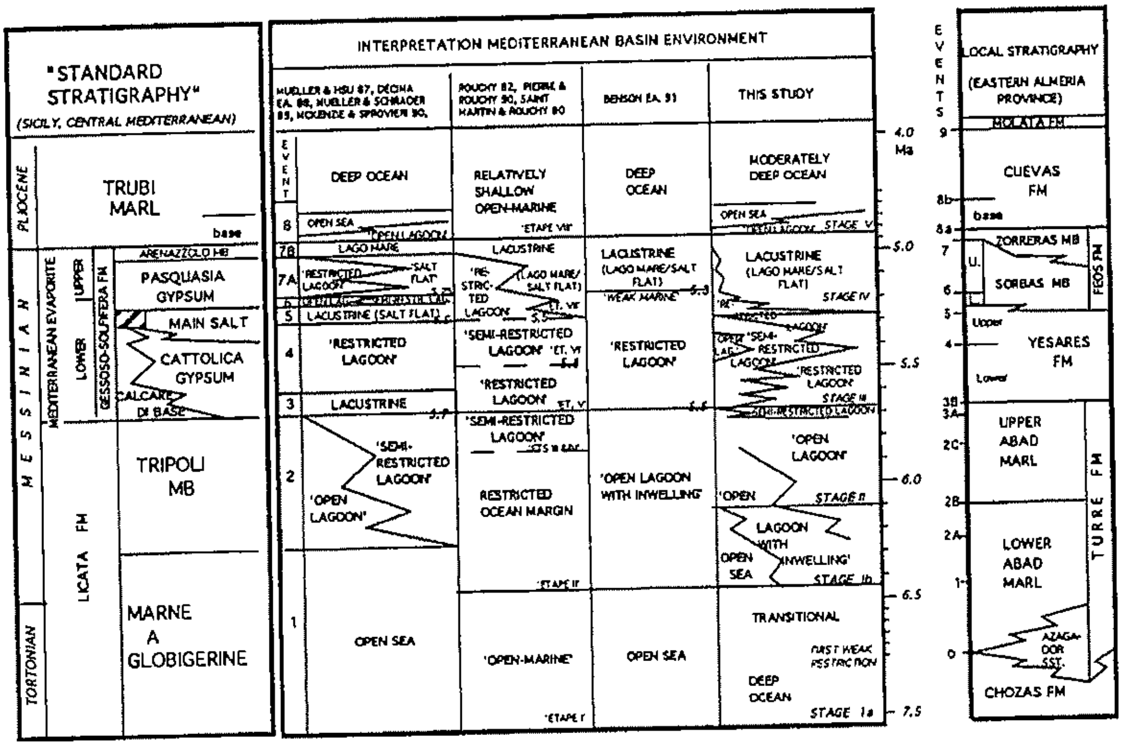


FIGURE 1.3. 'Standard stratigraphy' for the Mio-Pliocene of the central part of the Mediterranean Basin (to the left), its environmental interpretations in proposed 'scenarios' for the Mediterranean Messinian Event and those based upon this study (right)[1].

Explanation of marginal basin terminology in Fig. 1.2 Numbers refer to main facies events of Müller & Hsü (left) or to major facies boundaries (this study).

model, in which the earlier evaporites are essentially deposited under continuous marine inflow and in still relatively deep water, whereas more important isolation with lowered waterlevel within the Mediterranean Basin would have occurred during a second phase. However, no agreement has been reached on important 'details' such as the relative degree and exact timing of Mediterranean isolation. This becomes especially apparent from the divergence of opinions on a number of 'subevents' (Fig. 1.3): (1) environmental conditions during late early Messinian 'Tripoli' deposition are presently commonly interpreted as representing some degree of restriction in a partially enclosed marginal basin, but this degree is still relatively minor, already important, or extremely variable according to different authors (McKenzie et al. 1980,

[1]Numerical ages relative to the time-scale of Berggren et al. 1985. Recent publications (Harland et al. 1990, Hilgen 1991, Beets 1992, Vai et al. 1993, pers. comm. Berggren 1994) suggest that these ages are ca 0,5 million years to young for the Messinian, but they are here retained to facilitate comparison with other scenarios for the Messinian and with eustatic cycle charts. The position of Messinian 'events' in respect to the paleomagnetic reversal scale as discussed in Chapter 3.4 remains unchanged in our present concept.

Orszag Sperber et al. 1980, Rouchy 1981, 1982, Van der Zwaan 1982, Müller & Hsü 1987, Müller & Schrader 1989, Busson 1990, Benson et al. 1991); (2) a proposed important dessication at the base of Mediterranean Evaporite deposition (Hsü et al. 1977, Cita et al. 1978, Ryan & Cita 1978, Müller & Hsü 1987's 'Event 3', Decima et al. 1988) is not recognized in other models (Rouchy 1982, Benson et al. 1991, Busson 1990); (3) a major erosion surface at the top of the Lower Evaporites represents the main dessication ('Event 5') for Cita et al. 1978, Van Hinte et al. 1980, Müller & Hsü (1987) Benson et al. (1991), a major regression but without dessication of the deepest basins for Rouchy (1982, fig. 9 VI) and a tectonic event in Sicily for Busson (1990); (4) more or less important marine transgressions at the base of, and during 'Upper Evaporite deposition' (Events '6' and '7a' of Müller & Hsü 1987, Müller & Schrader 1989, Busson 1990, Müller et al. 1990, Pierre & Rouchy 1990), are not at all recognized by others , who propose consistent continental 'lago mare' conditions (Ruggieri 1967, Ruggieri & Sprovieri 1976, Van Couvering et al. 1976, Benson et al. 1991). Apart from this, basinal depth at the end of the Messinian, is still estimated somewhat differently amongst authors (Fig. 3.3, Stanley et al. 1976, Hsü et al. 1977, Montadert et al. 1978, Ryan & Cita 1978, Van der Zwaan 1982 (see also discussion in Busson 1990).

Besides differences in interpretation of the facies and geodynamic development of the Mediterranean Basin, also the relative importance and timing of the underlying factors of facies development (tectonic processes in the 'Entrance Area' and climate) are estimated differently.

Uplift in the Entrance Area must have played an important role in the temporary isolation of the Mediterranean Basin during its Messinian Salinity Crisis (e.g. Ruggieri 1967, Selli 1973, Hsü et al. 1973, 1977, Van Couvering et al. 1976, Rouchy 1982, Müller & Hsü 1987, Weijermars 1988, Benson et al. 1991). The relative importance of latest Miocene marine sedimentation in this area (Völk & Rondeel 1964, Völk 1966, Perconig 1966, 1976, *et al.* 1977, Montenat 1973b, Montenat et al. 1976, 1980, Müller & Hsü 1987, De la Chapelle 1987), obviously ports on the degree of Mediterranean isolation that could be reached by closing of the entrance. The Messinian episode is sometimes considered an interval rich in 'local' tectonic events, such as the subsequent closure of the 'North Betic' and 'Rif' straits (a.o. Benson et al. 1991), and wrench-fault tectonics in the eastern Betics (Montenat et al. 1990), whereas others have emphasized the importance of eustacy (Esteban & Giner 1980, Dabrio et al. 1981, Müller & Hsü 1987, Saint-Martin & Rouchy 1990, Pomar 1991).

Great or only relatively minor importance, lastly, is attributed to variations in global and/or Mediterranean climate as a factor in its facies development during the Messinian (Bizon et al. 1979, Dronkert et al. 1979, Chamley & Robert 1980, Troelstra et al. 1980, Van Gorsel & Troelstra 1981, Rouchy, 1982, Müller & Hsü 1987, Müller & Schrader 1989, Van de Poel 1989, Busson 1990, Suc & Bessais 1990, Van Hinte 1990, Benson et al. 1991).

THE MIO-PLIOCENE OF SE SPAIN

The first 'modern' stratigraphic work on the Neogene of Almeria, mainly centered upon the Vera and Sorbas Basins, dates back to Durand Delga & Magné (1959), Hermes (1962), Völk & Rondeel (1964), Ruegg (1964), Rondeel (1965), Völk (1966b, 1967), who established the local stratigraphic framework and the approximate age of the deposits. Further precisions and extension of the area to the south, were subsequently given by Jacquin (1970) and Montenat (1973a).

During the later seventies and early eighties more detailed work was published on the Messinian. Some papers dealt with its general stratigraphic aspects (Iaccarino et al. 1975, Montenat et al. 1976, 1980, Perconig 1976, Dronkert & Pagnier 1977, Dronkert et al. 1979, Van de Poel et al. 1984), whereas others primarily treated various aspects of its pre-evaporite basinal facies (Geerlings 1977, Gonzalez Donoso & Serrano 1977, Roux & Montenat 1977, Montenat et al. 1978, Civis et al. 1979a, Troelstra et al. 1979, 1980, Poore & Stone 1980), of its evaporites and associated facies (Rouchy 1976, Dronkert 1977a & b, 1978, Schleich 1977, Beets & Roep 1978, Montenat et al. 1980) and of its post-evaporite facies and the transition to the Pliocene (Montenat et al. 1976, Roep & Beets 1977, Gonzalez Donoso & Serrano 1978, Pagnier 1978, Civis et al. 1979b, Roep et al. 1979, Roep & Van Harten 1979, Cita et al., 1980, Clauzon 1980, Geerlings et al. 1980 a and b, Van de Poel et al. 1984). Still others dealt with the Messinian carbonate platform facies (Esteban et al. 1977, Pagnier 1977, Dabrio & Martin 1978, Armstrong et al. 1980, Esteban and Giner 1980, Dabrio et al. 1980). Special mention must be made of the work on the Messinian presented in the Ph.D. theses of Ott D'Estevou (1980) and Dronkert (1985).

Work on older Miocene and Pliocene deposits and on the 'neotectonics' of the area was presented in papers by Völk (1966b), Bousquet & Montenat (1974), Bousquet et al. (1975a & b), Bousquet & Philip (1975), Montenat & Ott D'Estevou (1977, 1987, 1990), Addicott et al. (1978), Harvey & Wells (1987), Coppier et al. (1990), Mather (1994) and in the Ph.D. theses of Ott D'Estevou (1980), Postma (1983), De la Chapelle (1987), Kleverlaan (1989) and Boorsma (1993).

Recent publications in which sedimentology, paleontology and (sequence-)stratigraphy of the Messinian deposits of the EAP are treated more or less extensively are from Müller (1986, 1993), De Deckker et al. (1988), Mankiewicz (1988), Van de Poel (1989, 1991, 1992), Saint-Martin & Rouchy (1990), Franseen & Mankiewicz (1991), Riding et al. (1991a & b), Jimenez & Braga (1993) and Martin et al. (1993).

Parallel to the study of the Central Mediterranean, the knowledge about Messinian facies and events in SE Spain has greatly improved over the years. However, also here the amount of isolation during deposition of certain facies units is still somewhat differently interpretated (e.g. Roep & Van Harten 1979, Montenat et al. 1980, Troelstra et al. 1980, Dabrio et al. 1981, Dronkert 1985, Müller & Hsü 1987, De Deckker et al. 1988, Riding et al. 1991, Van de Poel 1991, 1992, Jimenez & Braga 1993, Martin et al. 1993). In particular, interpretation of the exact age of the deposits and of the relative importance of unconformities is still rather confused (e.g. Fig. 1.4), which interferes with the interpretation of the local and regional environmental setting and with the distinction between local tectonic and global/Mediterranean eustatic and/or climatic nature of the 'events'.

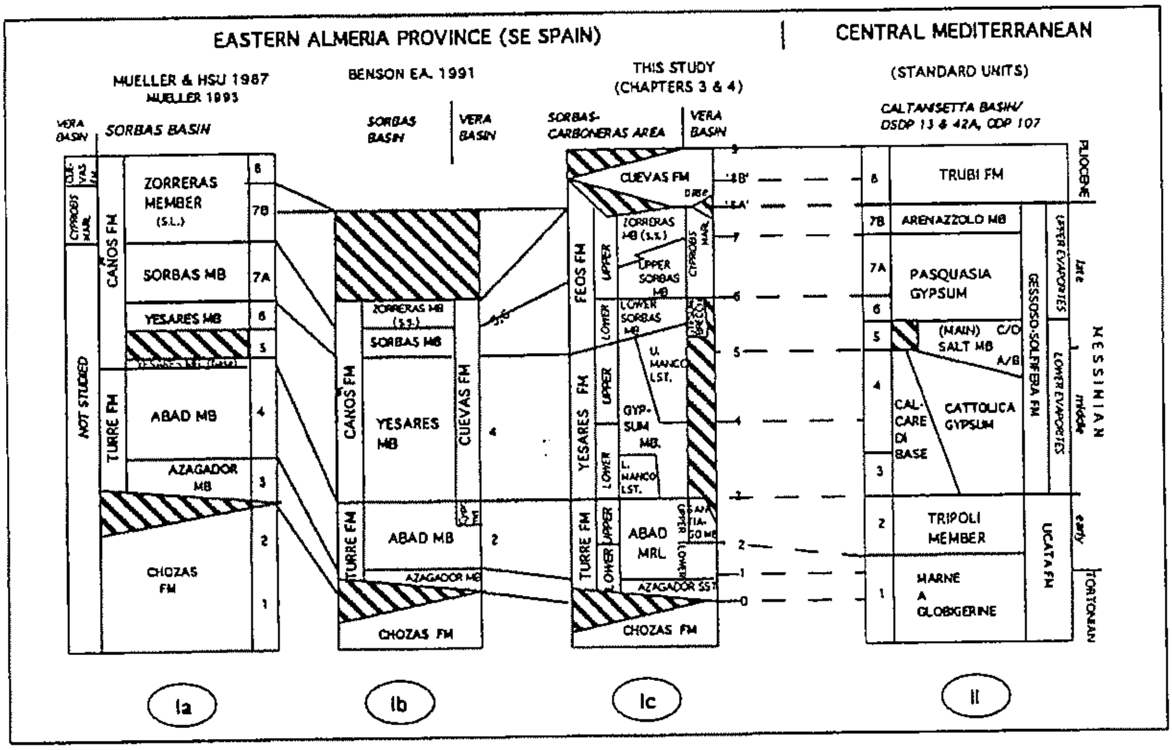


FIGURE 1.4. Different interpretations of the Mio-Pliocene stratigraphy of the deeper depositional areas of the Eastern Almeria Province (Ia-c) and of its correlation to the Central Mediterranean standard units (II). Numbers along columns refer to facies events ('Event stratigraphy' of Müller & Hsü 1987), or major facies boundaries (Ic).
Column Ia represents the 'classical' interpretation in which the Messinian evaporites and brackish-water deposits of the EAP ('Caños Formation') are underlain by an important regional hiatus and correlated to the late Messinian Upper Evaporites of the Central Mediterranean. Benson et al. (1991), on the contrary, propose the complete absence of late Messinian strata in the EAP and correlate the entire Caños Formation with the Lower Evaporites of the Central Mediterranean (Column Ib). In our interpretation (Ic), major basinal Messinian unconformities only occur in the coastal areas of the EAP (e.g. Vera Basin), whereas the Mio-Pliocene series is more or less complete in the more inland Sorbas and Northern Nijar basins. The 'lower' and 'upper Caños' (our Yesares and Feos Formations) then approximately correlate with Central Mediterranean Lower and Upper Evaporites, respectively. Beets & Roep (1978), Roep & Van Harten (1979) and Montenat et al. (1980, 1990) also suggested more or less continuous Messinian sedimentation in the central parts of the Sorbas Basin.

METHODS AND PRESENTATION

Facies and stratigraphic analysis are the primary tools used in this study. Environmental interpretation is based on the study of a number of facies aspects, such as the sediment type, its fossil content, in particular the benthic foraminifers, and the presence of authigenic minerals, as defined by field, microscopic and geochemical observations. In the basin analysis both classical and 'sequence stratigraphic' techniques were used.

Detailed sections were measured and lateral development of defined units mapped in the more central parts of the Northern Nijar Basin, where also some marginal sections were studied (Chapter 3). The stratigraphy of the other basins of the EAP (Sorbas, Vera and Agua Amarga) were studied for comparison, whereas intervals which are there better developed, or whose development is of particular interest, were studied in more detail (Chapter 4).

The facies analysis allows for interpretation of environmental conditions and paleogeographic setting of the studied area, whereas the stratigraphic analysis further adds to the understanding of these subjects and addresses the underlying factors (tectonic and climatic geodynamic processes) in the facies and basin development. For the distinction of the local from the regional and global signals, some field, laboratory and literature study was made of Central Mediterranean sections as well as of some areas in more western parts of the Betic-Rif orogen. Evaluation of the data in the light of the regional marginal basin development and a discussion on the relative importance of some underlying geodynamic processes are given in Chapter 5, whereas main conclusions are summarized in Chapter 6.

CHAPTER 2

REGIONAL AND LOCAL SETTING

The Neogene sediments of this investigation crop out in intra-montane and coastal basins of the Betic Cordilleras of southern Spain (Fig. 2.1). This mountain chain forms, together with the Rif mountains of northern Morocco, the westernmost part of the 'Arc of Gibraltar', which incompletely separates the Mediterranean Basin from the open ocean, leaving a connection at its central-western end with the Atlantic (Fig 1.1). Rupture, sedimentation and collision at the south Iberian and north African plate margins during the alpine cycle played a major role in its generation (e.g. Rehault et al. 1984, Hermes 1985, Sanz de Galdeano 1990, De Jong 1991).

The intra-montane and coastal basins of the Betic-Rif orogen to the north and south of the Gibraltar Strait form relatively low-lying areas between the Atlantic and Mediterranean domains. The large Guadalquivir trough (20 in Fig. 2.1) is the typical coastal Atlantic, 'external' basin of the Betic Cordilleras. The large intra-montane Baza-Guadix and Granada depressions (18-20) are referred to as 'central Betic' basins. Typical Mediterranean coastal, 'internal' basins underlie the plains of Malaga and Almeria, along the northern border of the Sea of Alboran, and are found along the central western Mediterranean (Balearic Basin) in the Alicante, Murcia and Almeria Provinces. The eastern part of the latter forms the major subject of this study. The present watershed is located relatively far to the east which makes that rivers of central and western Andalusia flow to the Atlantic, whereas only the sparse run-off in the eastern Betic Cordilleras (Dronkert 1985) is directed towards the Mediterranean (Fig. 2.1).

ENVIRONMENTAL SETTING OF THE MEDITERRANEAN BASIN

The present Mediterranean Sea can be seen as a remnant of the former Tethys Ocean which separated the Eurasian and southern continents during earlier stages of the alpine cycle (Benson et al. 1991). However, new oceanic basin formation has also taken place (Kastens, Mascle et al. 1990, Sanz de Galdeano 1990, Coutelle 1992). The present Mediterranean Basin therefore has some oceanic properties (presence of oceanic crust and waterdepths commonly over 2000 m), but its water mass is 'non-oceanic' (e.g. Fig. 1.2.Ic). As a result of the plate movements, the Mediterranean Basin became more and more isolated from the world-ocean during the Miocene. Separation from the Indian Ocean to the east probably took place during the Early Miocene (Van Couvering 1972).

The water-mass of the Mediterranean Basin has a relatively temperate and uniform temperature, relatively high and uniform salinity, and a low nutrient content (e.g. Sverdrup et al. 1942, Brongersma Sanders 1970). The elevated salinity and the absence of distinct thermal layering (deep waters are 12°C and no typical 'psychrosphere' is thus developed) are properties that differ from those of the open ocean. These properties are the result of a combination of the semi-enclosed nature of the basin, its sill depth and the regional climate with a negative water-balance (dominance of evaporation over precipitation and run-off), leading to an anti-estuarine water-exchange. The shallow inflow and deep outflow over the Gibraltar Sill (ca. 300 m) is very effective and relatively rapid (residence time of Atlantic water in the Mediterranean is only ca. 70 yrs: Benson et al. 1991). The present Mediterranean can be considered representative of a giant marginal basin of 'open sea' type, with only relatively minor restriction of the bioenvironment resulting from the effectivity of the circulation system (Fig. 1.2.Ic).

The relatively warm and saline 'Mediterranean Outflow Water' has an important influence upon circulation in the Atlantic Ocean and former changes in its volume may have influenced the climate in NW Europe (e.g. Blanc & Duplessey 1982) or even globally (Ryan 1973, Van Hinte 1982, 1990, 1992). The exact influence is however still a matter of discussion (e.g. Thunnell et al. 1987).

The present situation contrasts with Pliocene times, when an oceanic-type watermass with more distinct thermal layering, higher nutrient values and probably an 'estuarine circulation pattern' existed (Van Harten 1983, Thunnell et al. 1987, Hasegawa et al. 1990). For the Late Miocene a situation comparable to the present is commonly proposed (e.g. Vergnaud Grazzini 1983, Müller & Hsü 1987, Benson et al. 1991), with an alternation of anti-estuarine ('lagoonal') and estuarine circulation during the Messinian (Van Gorsel & Troelstra 1981, Müller & Hsü 1987, Müller & Schrader 1989). Van Hinte (1990, 1992) underlined the importance of variation between the 'anti-estuarine' and 'estuarine mode' of the Mediterranean Basin. The geodynamic development of the Mediterranean Entrance Area (MEA) in combination with changes in global and regional climate are at the root of different water-mass properties of the Mediterranean in the geologic past.

PALEOGEOGRAPHY OF THE ENTRANCE AREA

During Miocene times, important water-exchange between the Mediterranean and Atlantic took place through the so-called North-Betic and South-Rif Straits. These seaways, traversing the 'Gibraltar Arc Sill System' ('GASS'), were first recognized by French workers (Gentil 1916, 1918, Gignoux 1934) and became recently generally acknowledged (Ruggieri 1967, Hsü et al. 1973a, 1977, Benson 1976, etc.). Other Miocene waterways existed in more southern parts of the Betic Cordilleras (Montenat 1973a, 1977, Van Couvering et al. 1976, Rodriguez Fernandez 1982, Guerra- Merchan & Serrano 1993). Relative importance and time of closure of these different seaways is still somewhat open to discussion, but all data agree on their closure before the beginning of the Pliocene (see also hereafter).

A Pliocene marine connection must then have existed in the area of the present Strait of Gibraltar, but no agreement exists on the time of its opening, which has been proposed for different moments in the Miocene already by various authors (Ruggieri 1967, Hsü et al. 1973a, Montenat et al. 1975, Weijermars 1988). Some present data which are most pertinent to the history of the marine connections through the Gibraltar Arc in my opinion are summarized hereafter.

Data from diverse basinal areas in the central-northern Betics (e.g. 14, 20, 25, 26 of Fig. 2.1), indicate that the 'North Betic Strait' closed before the end of the Tortonian (Tjalsma 1971, Montenat 1973a, 1977, Campo Viguri 1977, Perconig & Fresneda 1977, Calvo et al. 1978, Ott D'Estevou et al. 1988, Servant-Vildary et al. 1990, unpublished data of J.J. Hermes, H. Dronkert, O.J. Simon and the author).

As for the Miocene seaways in the southern part of the Betic Cordilleras, the one through the 'Velez Rubio Corridor' (16), connecting Baza and Lorca basins, and the one traversing this area via Granada Basin and 'Alpujarran Corridor' (9) probably also closed during the Tortonian, or even before its beginning (Geel 1977, 1978, Rodriguez Fernandez 1982, Rodriguez Fernandez et al. 1990). Two of these 'South Betic Straits', however, one connecting the western Guadalquivir Basin with the basin of Malaga (through 22 and/or 23), and one connecting this area with the eastern Betics via the Basin of Guadix (18) and the 'Almanzora Corridor' (13), probably remained open up to a relatively late moment in the Miocene (Montenat 1973a, 1977, Weijermars 1988, Servant-Vildary et al. 1990, Jimenez et al. 1991, Guerra-Merchan & Serrano 1993). Latest Tortonian-earliest Messinian transgressive series in the area of Cordoba (Fig. 1.2; Verdenius 1971, partially reinterpreted) and the presence of early Messinian marine sediments in the Guadix Basin and the area of Guadalahortuna to its NW (Molino, in Peña 1975, Rodriguez Fernandez

1982) suggest a connection during at least some part of the Messinian. The presence of typical late early Messinian buliminacean faunas in the youngest marine sediments exposed to the west of Guadix (La Peza area; Rodriguez Fernandez 1982, personal observations) however imply a difficult connection with the open marine environment over a sill with its axis somewhere to the west of this locality.

The east-west thinning of Tortonian sediments in the northern part of the Alboran Basin and the inverse relationship for the Pliocene (Gaibar-Puertas 1975) as well as the presence of restricted marine Messinian facies in a drilling to the west of the present Strait of Gibraltar (Vergnaud Grazzini & Pastouret 1980) suggest that the latter did not come into existence before the beginning of the Pliocene.

Gignoux (1937) already suggested that the youngest Mediterranean Entrance in Miocene times persisted in the southern Rif area. This 'Rifian Corridor' would have become closed at the end of the middle Messinian (Carbonel et al. 1981b, Benson et al. 1991). Data from Bernini et al. (1992) however indicate a relatively important sill in the early Messinian in this corridor also and a difficult connection through this area in younger Messinian times.

FIGURE 2.1. A. Geologic map of the northern-central part of the 'Mediterranean Entrance Area' showing the main tectono-stratigraphic features of the Betic Cordilleras and the location of the study area (encadrated). Slightly modified after Van de Poel 1992, fault-pattern mainly after Montenat et al. (1990), Sanz de Galdeano (1990a).
Pre-Neogene and Neogene-Quaternary rocks constitute numerous smaller mountain chains (Sierras) and low-lying areas, respectively. The basins of the Eastern Almeria Province, are located at the eastern end of E-W directed 'South Betic Corridors', which connect the basins in the central part of the Betic Cordilleras with the Mediterranean coast. The main basinal areas which further characterize these corridors are (from south to north): the Southern Nijar or Almeria (5), Dalias (6), and Malaga Basin (7) , the Upper Andarax or Tabernas (8), and Ugijar and Orgiva basins or Alpujarran Corridor (9), the San Miguel de Salinas (10), Puerto Lumbreras (11), and Albox Basin (12) and the 'Almanzora Corridor' (13) and the Fortuna (14), and Lorca (15) Basin and the Velez Rubio Corridor (16). These basins and corridors show numerous smaller interconnections and commonly contain at least up to middle Tortonian marine deposits, indicating the importance of marine connections in this area during the Miocene.
The Baza-Guadix (17-18) and Granada (19) basins are large intra-montane depressions of the central part of the Betic Cordilleras. Further to the north, two large 'foreland' basins, the 'Atlantic' Guadalquivir (20) and the 'continental' Albacete Basin (21) can be recognized. Their southern margins connect with the northern part of the large central depressions through numerous smaller pull-apart basins of the Alicante-Cadix (or 'Crevillente') Fault System, all characterized by 'older Neogene' marine deposits of the 'North Betic Strait'.
B. Location of Fig. 2.2.

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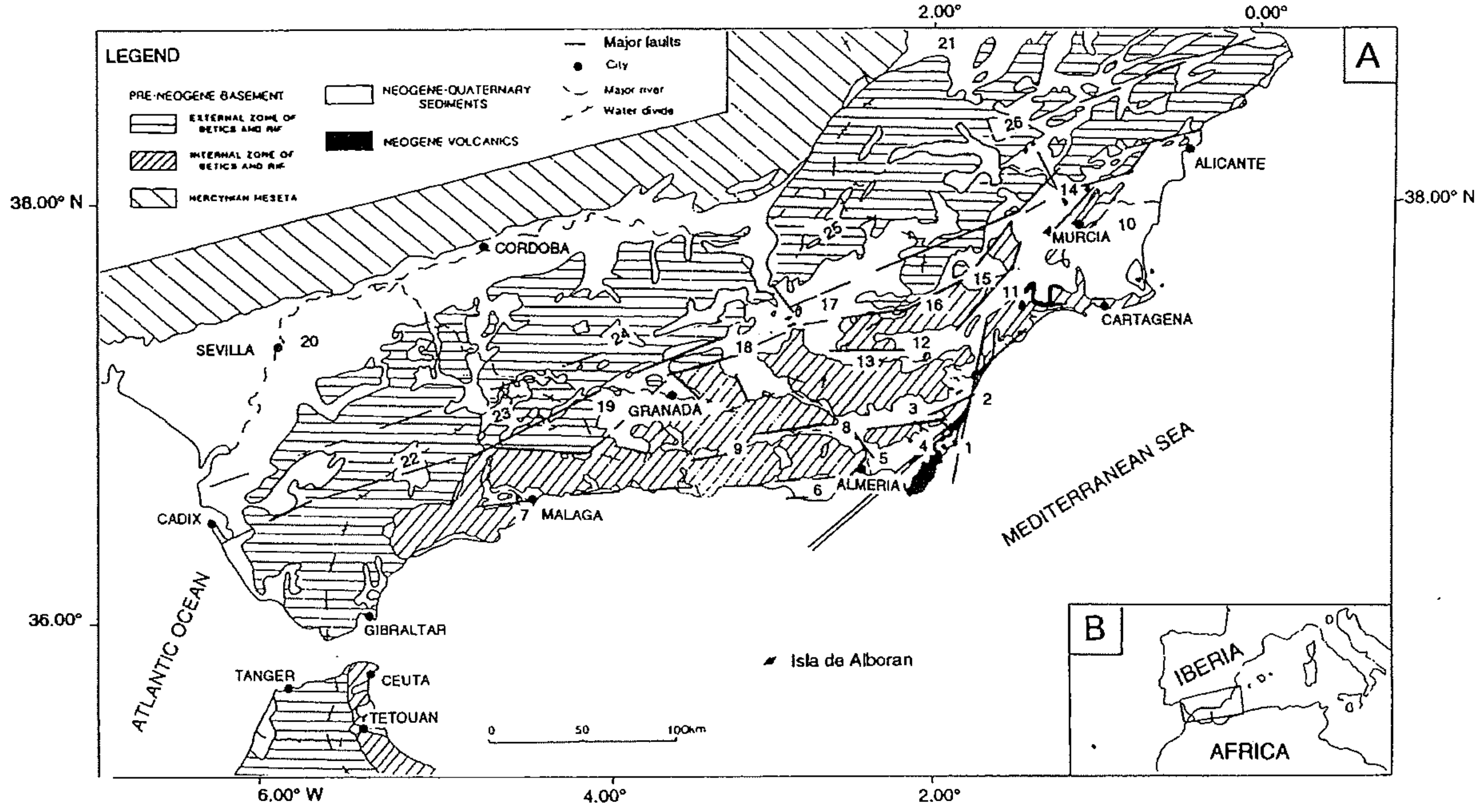
A
LEGEND
PRE-NEOGENE BASEMENT
EXTERNAL ZONE OF BETICS AND RIF
INTERNAL ZONE OF BETICS AND RIF
HERCYNIAN MESETA
NEOGENE-QUATERNARY SEDIMENTS
NEOGENE VOLCANICS
Major faults
City
Major river
Water divide
CORDOBA
SEVILLA
CADIX
GIBRALTAR
TANGER
CEUTA
TETOUAN
MALAGA
GRANADA
ALMERIA
CARTAGENA
MURCIA
ALICANTE
ATLANTIC OCEAN
MEDITERRANEAN SEA
Isla de Alboran
0 50 100km
38.00° N
36.00°
6.00° W
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1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20 21 22 23 24 25 26
B
IBERIA
AFRICA

STRUCTURAL SETTING OF THE BETIC NEOGENE BASINS

The Betic Cordilleras are subdivided in an Internal and an External Zone on basis of differences in stratigraphy and degree of metamorphism and tectonization of the rocks of its Sierras (Egeler & Simon 1969; Fig. 2.1). Those of the Internal Zone display a stack of nappes of different tectonic and metamorphic history, which builts the basement to the Neogene-Quaternary deposits of the intermediate plains. (Egeler & Simon 1969, De Jong 1991). In contrast to the External Zone, paleozoic rocks crop out abundantly.

From the latest Burdigalian onwards typical 'Neogene basin characteristics' existed in the Betic Cordilleras, since the main tangential movements were completed before the end of the Early Miocene (Hermes 1977, 1985, De Smet 1984, Sanz de Galdeano 1990a, Montenat et al. 1990, De Jong 1991). However, considerable tectonic activity continued up to at least the late Tortonian and westward movement of an 'Alboran Block may have continued during the Middle and Late Miocene (Leblanc & Olivier 1984, Sanz de Galdeano 1990a).

After first suggestions as to its activity (Hetzel 1919, Fernex 1964, Völk 1966b, Paquet 1969), the importance of wrench-fault tectonics in the origin, location, shape, facies development and tectonic style of the Neogene basins in the Betic Cordilleras has been well-documented (Bousquet & Montenat 1974, Bousquet et al. 1975a, b, 1977, Bousquet & Philip 1976, Armijo et al. 1977, Hermes 1978, 1985, Sanz de Galdeano 1983, *et al.* 1985, 1990, De Smet 1984a & b, Montenat et al. 1987, 1990, Martin Algarra et al. 1988, Bon et al. 1989, Van der Straaten 1990, Boorsma 1993).

Three major fault directions are recognized (Fig. 2.1). The first one, approximately E-W, often N70E comprises the Alicante-Guadix (or 'Crevillente') Fault, the 'Betic Fault System' of Hermes (1985), and the 'Almanzora', 'Alpujarran', and a third important fault which forms the boundary of the internal Betics with the Sea of Alboran. A second important fault trend is approximately NW-SE, commonly connecting basins in the E-W directed trends described above. Good examples are the the Tijar Fault Zone, that connects the eastern end of the Guadalquivir Basin with the Guadix-Baza Depression (Sanz de Galdeano 1983, personal observations) and the Andarax Fault, between the Tabernas and Southern Nijar (or Almeria) Basin (Postma 1983, Kleverlaan 1989). Both E-W and NW-SE trending faults are probably inherited from the early and middle alpine history of the Betic Cordilleras (Figs. 6.1-6.3 in De Jong 1991). A third important fault trend has a direction from N45E to still smaller angles and is most obvious in the eastern part of the Betic Cordilleras as the Eastern Betic Shear Zone of Montenat et al. (1987) or the Trans-

Alboran Shear Zone of De Larouzière et al. (1988).

During the Neogene-Quaternary, the relative activity of the different fault-trends has not remained constant. The approximately E-W faults were particularly active before the late Tortonian, whereas the NE-SW faults had their greatest activity in the youngest Neogene (cf. e.g. difference in disposition at 'Hueli-Polopos' and Serrata Fault Zones in Fig. 2.3). This has been explained by changes in the dominant compressional direction from NW-SE during the main part of the Miocene, to NNW-SSE in the latest Miocene and Pliocene (Ott D'Estevou & Montenat 1985, Montenat et al. 1987, 1990, Sanz de Galdeano 1987, 1990a & b, Boorsma 1993).

STRUCTURAL AND STRATIGRAPHIC SETTING IN THE EASTERN ALMERIA PROVINCE

As in other parts of the Betic Cordilleras, three main structural trends can be recognized in distribution and shape of Sierras and of Neogene Basins and interconnecting 'corridors' in the Eastern Almeria Province (Fig.2.2).

The approximately **E-W structural trend** is best recognizable in the shape of the Sorbas Basin and *'Lucainena'* and *'Los Gallardos'* corridors to its west and east, which can be considered a large approximately E-W trending asymmetrical syncline, and in the alignment of the Sierra de los Filabres which borders them to the north. The Sierras Alhamilla and Cabrera with their intermediate area (*'Polopos Ridge'*) form an irregular E-W trending anticline separating the Sorbas and Vera basins from the Nijar Basin to the south (Fig. 2.3). Smaller structural phenomena belonging to this typical Betic (or alpine) trend are the complicated tectonic zones, which show imbrication of basement and older Neogene material as well as imbrication of different basement units at the northern and southern borders of the Vera Basin (Simon 1963, Rondeel 1964, Völk 1967).

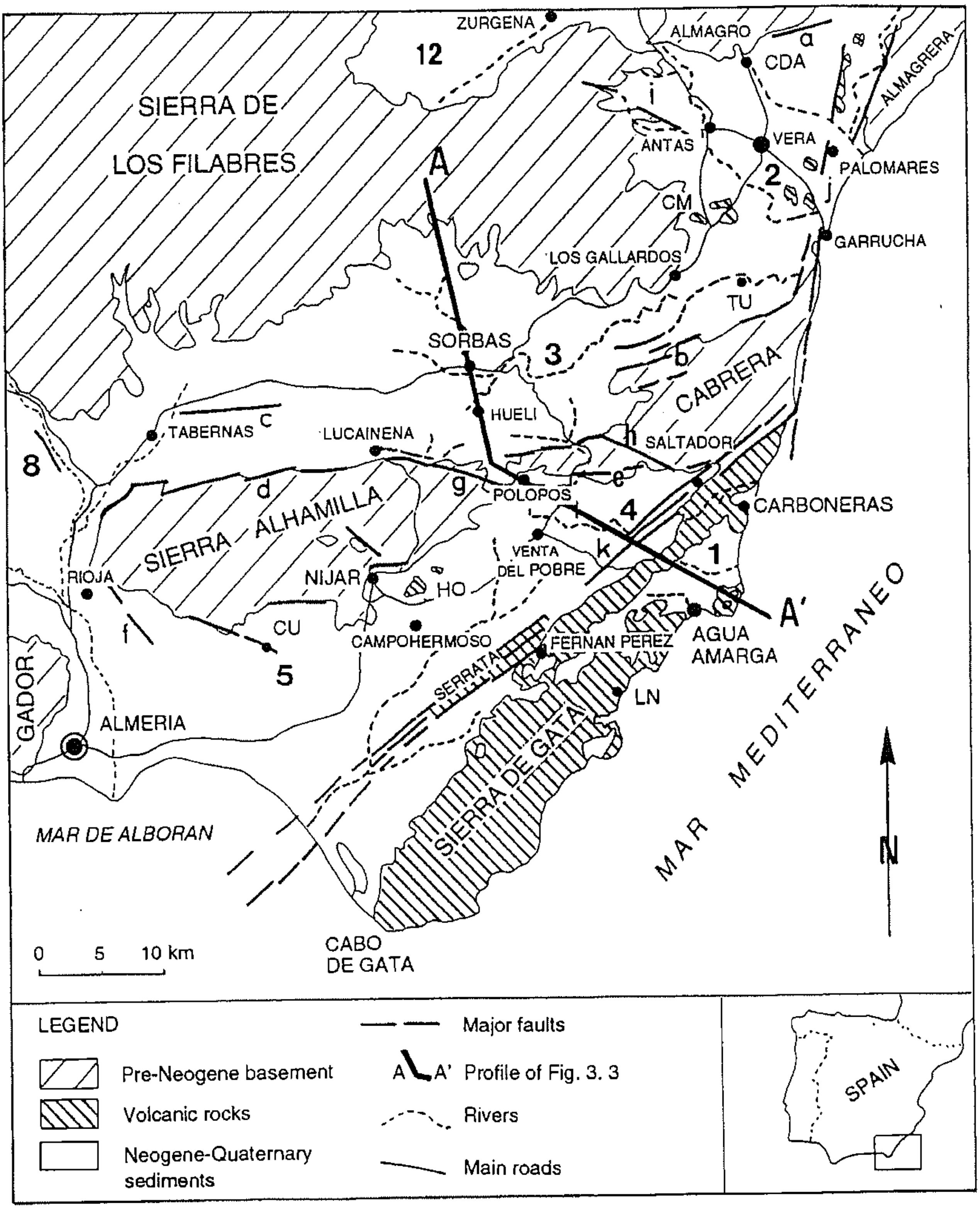
The **NW-SE structural trend** is most obvious in the area of Almeria where numerous NW directed faults are present of which the most important is the Andarax Fault (f in Fig. 2.2), which plays an important role in the *'Rioja Corridor'* (Postma 1983, Kleverlaan 1989), between the Southern Nijar and Gador-Tabernas basins. WNW directed structures are further present in the *'Feos Corridor'*, between the Sorbas and Northern Nijar Basins. Here, the most apparent structural phenomenon is the Gafarillos Fault (h; Rondeel 1965, Ott D'Estevou et al. 1990; Fig. 2.2). A parallel structure at its western end (Lucainena Fault of Estevou et al.

1990) is responsable for brecciation of Messinian gypsum with overlying Pliocene deposits in the area SE of Polopos. This structure may further be an element in defining the boundary and waterdivide between the Northern and Southern Nijar Basins (*'Venta del Pobre-Fernan Perez Sill'*) and the southern boundary of the Agua Amarga Basin. The latter basin may be viewed as another NW-SE corridor connecting the Northern Nijar Basin with the Western Mediterranean. The importance of the NW trend in basin connections is further demonstrated by NW trends of basement units and faults in Neogene-Quaternary rocks in the relatively low area *('Zurgena Corridor)* between the Sierra de los Filabres and the Sierra Almagro (Völk 1967, De Jong 1991). which connects the Vera with the Albox Basin to its NW (Fig. 2.2).

The **NE to N directed structural trend** is typically well-developed in the Almeria Province, in particular in its eastern part. It is most obvious in the shape of the Vera Basin and '*Serrata*' and '*Saltador*' corridors, and in the alignment of Sierra Almagrera, smaller basement blocks in the area of Garrucha, '*Castillico Ridge*' (Van de Poel et al. 1984) and Sierra de Gata, which border these depressions to the east (Figs 2.1 & 2.2).

These large structures are related to the *Palomares* and *Serrata (or Carboneras) Wrench Zones* along the eastern margins of the Vera and Nijar basins, that had sinistral lateral movements in late Neogene times . Several tens of kilometres of relative lateral displacement are most easily demonstrated for the Palomares Fault Zone (Völk 1967, Bousquet & Montenat 1974, Montenat et al. 1987, Bon et al. 1988, Barragan et al. 1990). Some lateral displacement along the Serrata Fault is most obvious in stronger tectonization of Messinian to Recent units in the proximity of

FIGURE 2.2. Map showing Neogene basins and main morpho-structural feautures of the Eastern Almeria Province (EAP). Numbers 1, 2, 3 and 4 refer to the Neogene basins of the EAP (Agua Amarga, Vera, Sorbas and Northern Nijar basins, respectively). Numbers 5, 8 and 12 refer to the adjacent Southern Nijar or Almeria, Gador-Tabernas and Albox basins, respectively.
The Neogene basins are more or less separated by small mountain chains (Sierras) consisting of pre-Neogene material or Miocene volcanics (between the Nijar and Agua Amarga basins). The basins are interconnected by a number of 'corridors'. The 'Rioja' and 'Serrata' corridors connect the Southern Nijar, with the Tabernas and N. Nijar basins, respectively. The 'Agua Amarga', 'Saltador' and 'Feos' corridors connect the N. Nijar Basin with the Mediterranean and the Sorbas Basin, the 'Lucainena' and 'Los Gaillardos' corridors, the Sorbas Basin with the Tabernas and Vera basins and the 'Garrucha', 'Pulpi' and 'Zurgena' corridors, the Vera Basin with the Mediterranean and the Albox-Huercal Overa Basin. Later, transpressive sills' nowadays occur within these corridors (e.g. 'Hueli-Polopos Ridge' in the Feos, 'Castillico Ridge' in the Agua Amarga and 'Antas Ridge' in the Zurgena Corridor) (see also Fig. 2.3).
Some other localities mentioned in the text: CdA = Cuevas del Almanzora; CM = Cabezo Maria; Tu= Turre; Ho = Hoyazo; CU = Cuevas Ubedas; LN = Las Negras.
a= 'S. Almagro Fault', b= N. Cabrera Fault, c= Marchante Fault, d= N. Alhamilla Fault, e= S. Cabrera Fault, f = Andarax Fault, g= Lucainena Fault, h= Gafarillos Fault, i= unnamed fault near Antas, j= Palomares, k= Serrata Fault Zone. -------------------->


ZURGENA
12
ALMAGRO
a
CDA
ALMAGRERA
SIERRA DE
LOS FILABRES
A
ANTAS
VERA
PALOMARES
2
CM
GARRUCHA
LOS GALLARDOS
TU
SORBAS
3
b
CABRERA
HUELI
TABERNAS
c
LUCAINENA
h
SALTADOR
8
d
g
POLOPOS
e
CARBONERAS
SIERRA ALHAMILLA
4
VENTA
DEL POBRE
k
1
RIOJA
NIJAR
HO
A'
CU
f
CAMPOHERMOSO
FERNAN PEREZ
AGUA
AMARGA
5
SERRATA
LN
GADOR
ALMERIA
SIERRA DE GATA
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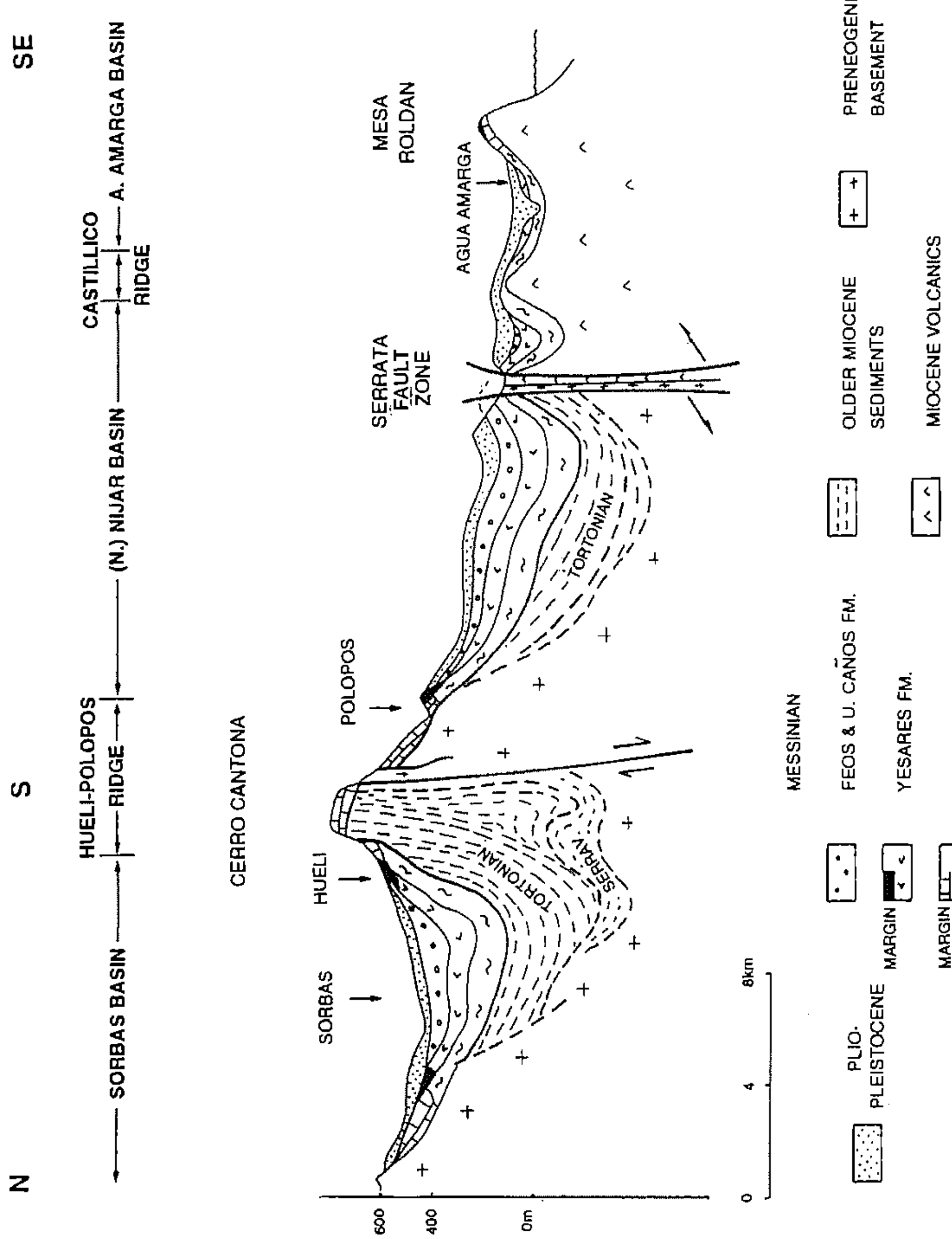
N
0 5 10 km
CABO
DE GATA
LEGEND
Pre-Neogene basement
Volcanic rocks
Neogene-Quaternary
sediments
Major faults
A A' Profile of Fig. 3. 3
Rivers
Main roads
SPAIN


N
S
SE
SORBAS BASIN
HUELI-POLOPOS RIDGE
(N.) NIJAR BASIN
CASTILLICO RIDGE
A. AMARGA BASIN
CERRO CANTONA
SORBAS
HUELI
POLOPOS
SERRATA FAULT ZONE
AGUA AMARGA
MESA ROLDAN
TORTONIAN
SERRAV.
TORTONIAN
600
400
0m
0
4
8km
PLIO-PLEISTOCENE
MESSINIAN
FEOS & U. CAÑOS FM.
MARGIN
YESARES FM.
MARGIN
TURRE FM.
OLDER MIOCENE SEDIMENTS
MIOCENE VOLCANICS
PRENEOGENE BASEMENT

this structure, horizontal striations on fault surfaces and in the sedimentary facies and synsedimentary deformation of Early Messinian deposits (Bousquet et al. 1975a, Greene et al. 1977, Montenat & Estevou 1990, Coppier et al. 1990, Boorsma 1993). The NW curvature of latest Tortonian to earliest Messinian sediments in the Saltador area indicates post-earliest Messinian movements of at least a few kilometers, whereas differences in stratigraphy between the area due west and east of the Serrata Zone in the El Argamason area may indicate displacement of at least a similar amount after the middle Pliocene (fig. 1 in Van de Poel 1991). The western borders of Vera and Northern Nijar basins and Serrata Corridor probably owe their existance to satellite faults of the Palomares and Serrata systems, marked by the isolated Late Miocene Cabezo Maria (CM) and Hoyazo (Ho) volcanic eruption centres (Figs 2.2 & 5.8). NE directed faults are further important in the connection area of the Sorbas and Vera Basins (Rondeel 1965, Los Gallardos Strait of Barragan et al. 1990).

The Agua Amarga and Vera basins are more or less open towards the Mediterranean (Balearic Subbasin) in the east, but the common exposure of Messinian and older rock-units near the shoreline and a relatively elevated position of the Pliocene deposits gives them the aspect of 'hanging basins'. The present eastern coast is probably largely defined by parallel faults belonging to the Palomares Fault System (e.g. Coppier et al., 1990). It has been suggested that the Agua Amarga and Vera areas were more closed off from the Mediterranean during Neogene times by presently foundered volcanic and pre-Neogene landmasses to their east (Völk 1966b, Montenat et al. 1976, Bellon et al. 1983, Coppier et al. 1990). The most typical example of an 'open end' is the southern part of the Nijar Basin, which slopes relatively gently into the Gulf of Almeria and hence in the sea of Alboran.

The Neogene basins of the EAP have an early Burdigalian to Recent sedimentary fill which shows an uneven disposition, both in space and in time (Fig. 2.3). The basins often show an asymmetrical cross section with a 'stable' margin to the NW, which shows late Tortonian sediments directly transgressive upon the pre-Neogene basement, followed by a relatively well-preserved but thin, shallow-marine Mio-Pliocene series. The opposite margin commonly has a more 'active' aspect in the presence of Messinian deeper-water deposits and exposure of Tortonian and older Neogene rocks and in the presence of some strong interformational un-

<----- **FIGURE 2.3.** Simplified geologic cross-section through the Northern Nijar and adjacent basins, showing their sedimentary fill and main structural features (Fig. 2.2 for location).
Note the relative importance of young, differential uplift reflected in the present elevations of the, approximately contemporaneous, shallow marine Pliocene deposits (stippled).

conformities (Figs 2.3, 5.7-5.9). The central part of the basins has a more or less thick Late Tortonian to Quaternary series.

The small Agua Amarga depression forms a particular case, since it is encased in the SE margin of the Northern Nijar Basin, which is predominantly constituted by pre-Messinian Neogene volcanic rocks. Here, the stable and more active margins are situated in the SW and NE, respectively. Similar disposure of Neogene rock units is found in the Rioja, Feos and Zurgena Corridors.

The oldest Miocene sediments are only exposed as tectonic slivers in the S. Almagro, N. Cabrera, and Serrata Fault Zones (Völk & Rondeel 1964, Van de Poel 1980, in Pineda Velasco et al. 1983, 1991, Boorsma 1993; Fig. 2.2). Serravallian sediments have a somewhat greater distribution both in the above mentioned fault zones, in the N. Alhamilla and Gafarillos Faults at the northern and eastern borders of the Polopos Ridge (Rondeel 1965, Volk 1967, Ott D'Estevou 1980), and along the southern margin of the Sierra Alhamilla from Nijar westward (data from Ochoa et al. 1973a and H. Dronkert, J.A. Manuputty, G. Postma and the author unpublished). Tortonian sediments are again better preserved and are found both, as slivers in the N to NE directed Palomares and Serrata fault zones and as a virtually continuous exposures at the 'active' southern margins of the Vera, Sorbas and Tabernas basins and the E margin of Rioja Corridor. Good exposures of Messinian deposits are, apart from their development along the 'stable' northern to western margins of the Sorbas, Nijar and Vera Basins and the southern Agua Amarga Basin margin, also found basinward from the margins of these areas (Fig. 2.3).

In the central parts of the basins close to the present coastline (Vera, Agua Amarga and Southern Nijar basins) relatively thick marine Pliocene sections are exposed. Towards more central parts of the Betic Cordilleras, on the other hand, Tortonian, respectively Serravallian deposits form the main exposed central fill in Gador-Tabernas Basin and 'Alpujarran Corridor', respectively (Kleverlaan 1989, Sanz de Galdeano et al. 1985, Rodriguez Fernandez et al. 1990).

CHAPTER 3

THE NORTHERN NIJAR BASIN

3.1. INTRODUCTION

Detailed study was made of the Mio-Pliocene sediments of the more central part of the Northern Nijar Basin. A first overview of the Messinian units recognized and mapped in this area (Rondeel 1965, Westra 1969, Van de Poel 1976, 1980, Mallee 1977), was given in Dronkert et al. (1979), Pineda Velasco et al. (1983) and Van de Poel et al. (1984).

Our first detailed paper on this basin (Van de Poel 1991) was centred on the stratigraphy, sediment-petrography and geochemistry of the 'gypsum-ghost limestones', which are common in the Messinian evaporite series of this area. The diagenetic history of this lithology is discussed, whereas also an environmental interpretation is given of the underlying, overlying and laterally equivalent Messinian deposits. Comparisons are made with the classical Messinian units of the central Mediterranean and a correlation-scheme proposed.

In a second paper (Van de Poel 1992) the foraminifer content of the Mio-Pliocene sediments of this area is described, together with some other fossil constituents and the authigenic mineral content of the wash-residues. Five successive fossil assemblages are discerned, primarily on basis of the smaller benthic foraminifers. The assemblages are compared with those described from modern to subrecent environments, both within and outside the Mediterranean Basin, and interpreted in terms of bottom-water properties (depth, salinity and oxygen-content). Important temporary changes in these properties during the Messinian, together with long-term changes over the whole Mio-Pliocene interval, allow for the recognition of five successive 'Environmental Stages', each of which is represented by its characteristic 'microfacies'. The main facies boundaries are dated, primarily by means of the planktonic foraminifer biostratigraphy and to some extent by calculation of sedimentation rates and correlation with well-dated events of other areas in the western-central Mediterranean. On basis of comparison with time-equivalent facies from more central parts of the Mediterranean and from basins on the Atlantic side of the Entrance Area, conclusions are drawn on paleogeography and effectivity of the GASS in the Mio-Pliocene boundary interval.

Finally, the Mio-Pliocene stratigraphic and facies development of the basin is described in terms of sequence stratigraphy (Van de Poel 1989). The proposed scheme is primarily based on the development in the (Northern) Nijar Basin, but additional data are used from the Sorbas, Vera and Agua Amarga basins and the scheme is applicable to the entire area.

In contrast to earlier and more recent papers on this subject, which are primarily based on the stratigraphy at the basin margins (Esteban & Giner 1980, Dabrio et al. 1981, Saint Martin & Rouchy 1990, Franseen & Mankiewicz 1991, Riding et al. 1991, Martin & Braga 1993), here the development in the central parts of the basins is taken as a starting point. The subdivision in sequences and systems tracts supplies a more genetic scheme for describing the stratigraphic and facies development of the Eastern Almeria Province and facilitates the evaluation of the relative importance of different factors (local tectonic and sedimentary processes, eustatics, climate) in its geodynamic development.

3.2. MESSINIAN STRATIGRAPHY OF THE NIJAR BASIN (S.E. SPAIN) AND THE ORIGIN OF ITS GYPSUM-GHOST LIMESTONES[1]

H.M. van de Poel

SUMMARY

The middle Messinian of the Nijar-Carboneras area (S.E. Spain) shows common rapid lateral transitions from thick-bedded massive gypsum to brecciated or massive limestone with voids and pseudomorphs after gypsum crystals. These 'gypsum-ghost limestones' are underlain by, and interbedded with, laminated marly sediments that contain a restricted marine microfauna attesting to oxygen-deficient conditions. Oolite-rich series of the basin margin, which include gypsiferous stromatolite and a few restricted marine fauna levels, probably constitute a lateral equivalent.

Upper Messinian fine-grained laminites of the central part of the basin contain brackish fossil assemblages and numerous tongues of coarse clastic material derived from the basin margins.

The gypsum-ghost limestones are interpreted to be essentially the product of two phases and types of diagenesis. Microbial sulfate-reduction during oxygen-deficient periods of the middle Messinian first played a role in their formation. An important fresh-water diagenetic phase took place later, probably in the late Messinian.

INTRODUCTION

Unfossiliferous limestone containing pseudomorphs after, and voids moulding, gypsum crystals, is common in the Messinian of Carboneras (Figs 1-3, Plates I-IV) (Van de Poel et a l 1984, Dronkert 1985). This 'gypsum-ghost limestone' was referred to as 'Calcare di Base-like rock' (Dronkert et al. 1979) because of its similarity in field-texture and stratigraphic position to the classical Messinian Calcare di Base of Sicily (Richter Bernburg 1973, Decima et al. 1988). This lithology also occurs in the 'caprock' of salt domes (Richter Bernburg op. cit. Pierre & Rouchy 1988).

The interest of this type of rock lies both in the position it takes in scenarios for the Messinian Salinity Crisis (Hsü et al. 1977, Müller & Hsü 1987), and in its reservoir potential for hydrocarbons and metal sulfides (Friedman 1980, Rouchy et al. 1985). Different mechanisms, however, have been proposed for its origin. The Sicilian Calcare di Base has been interpreted to be essentially either a primary carbonate deposit (Richter Bernburg 1973, Decima et al. 1988) or the product of microbial transformation of sulfate (Dessau et al. 1962; Neev and Emery 1967, Pierre & Rouchy 1988). Fresh-water diagensis of massive sulphates, proposed as origin for similar rocks from the Purbeck (uppermost Jurassic) of southern England by West (1973, 1975), is a mechanism, that has so far been little considered for the Messinian.

[1]Text and figures earlier published in *Geologie en Mijnbouw 70 (1991), pp. 215-234.*
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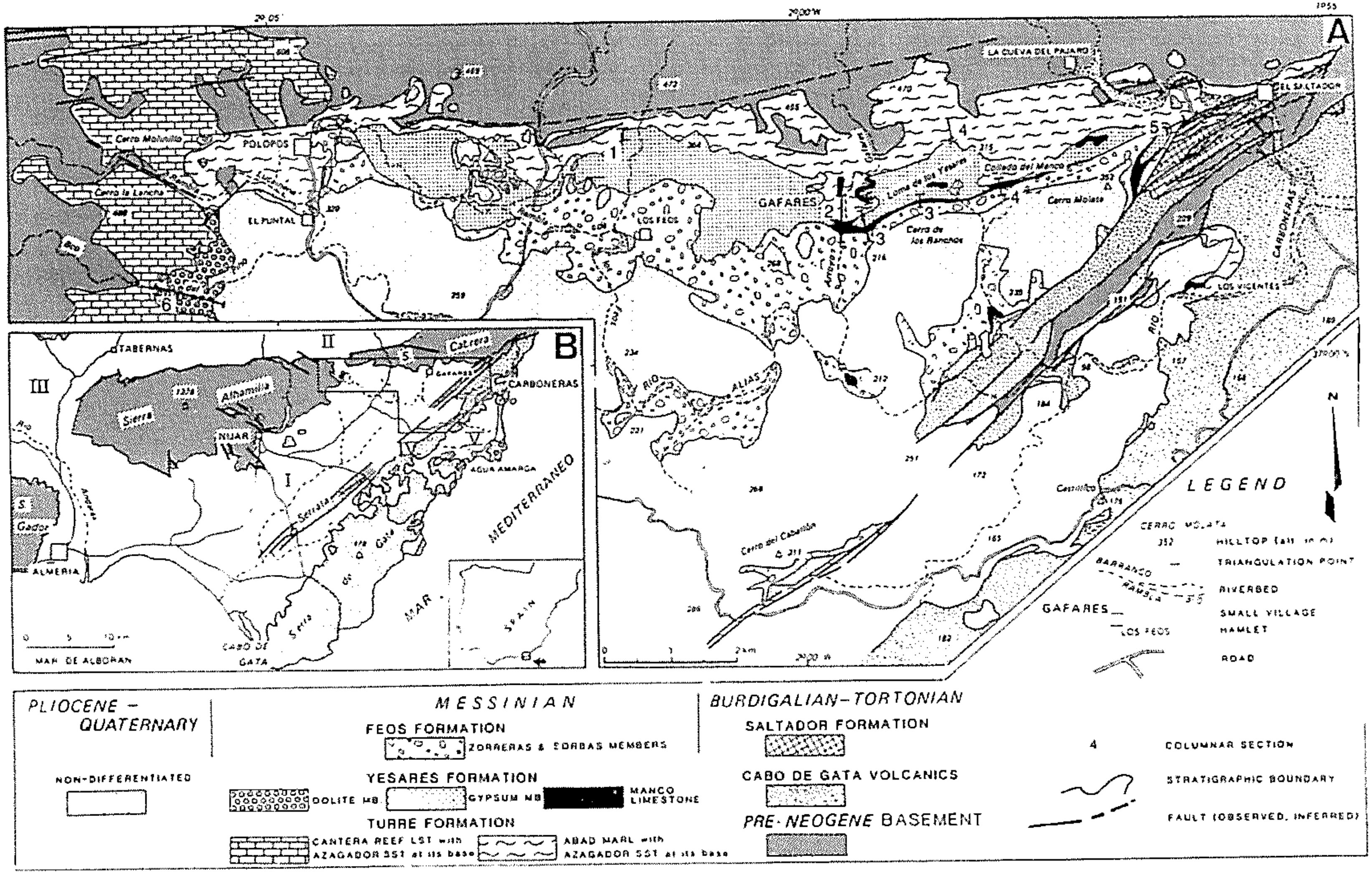


Fig. 1 Geological map of the Northern Nijar Basin (A), with distribution of gypsum ghost limestones (in black), locations of sections discussed in text and its regional setting (B) (I = Nijar Basin; II = Sorbas Basin; III = Tabernas Basin; IV = 'Castillico Ridge'; V = Agua Amarga Basin).

The objective of this chapter is to give a detailed description of facies and stratigraphy of the 'gypsum-ghost limestone' and associated sediments from the Northern Nijar Basin and to reconstruct the basin configuration and paleoenvironmental conditions that led to their formation.

GEOLOGICSETTING

The Nijar Basin is a vast SW-NE elongated depression near Cabo de Gata (Fig. 1B). It is one of the Neogene-Quaternary basins of southern Spain, which formed by pull-apart of the alpine basement of the Betic Cordilleras, and is situated in the 'Eastern Betic Transcurrent Shear Zone' of Montenat et al. (1987). The Nijar Basin is bounded to the north and west by intensely tectonized, and mostly metamorphic, pre-Neogene rocks of the Sierras Cabrera, Alhamilla and Gador, which are separated by lower areas connecting the basin to the Sorbas and Tabernas basins in the north (Fig. 1B). The basin is bounded to the southeast by the Miocene volcanics of the Sierra de Gata (Fuster et al. 1965, Bellon et al. 1976). A northern continuation, the 'Castillico Ridge', forms a hilly range separating the Northern Nijar Basin from the Basin of Agua Amarga (Van de Poel et al. 1984). The Castillico Ridge runs roughly parallel to the NW-SE trending left-lateral Serrata-Carboneras Wrench Zone, which still shows some activity today (Bousquet & Philip 1976, Greene et al. 1977). To the southwest, the Nijar Basin runs out into the Sea of Alboran of which it can be considered an uplifted part (Fig. 1B).

Most outcropping sediments of the Nijar Basin are marine Pliocene and continental Quaternary (Fig. 1A) (Van de Poel et al. 1984). Older, Messinian material, is found in areas directly adjacent to the Sierras and in the Serrata Wrench Zone, which, near the

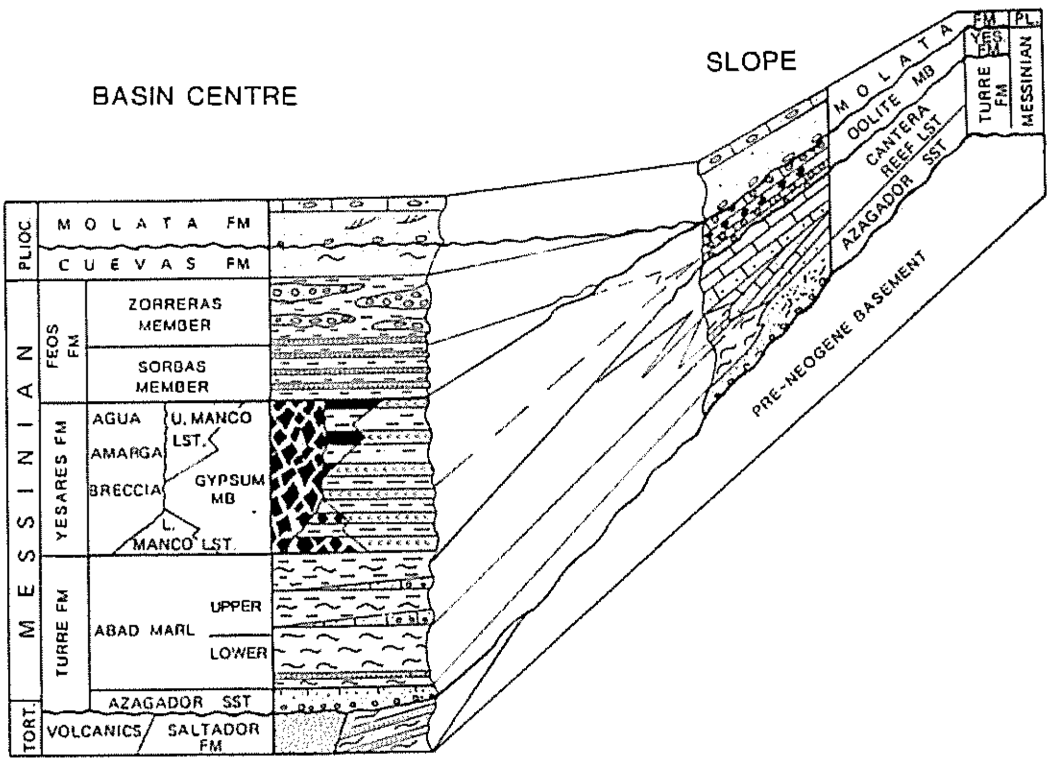


Fig. 2A. Lithostratigraphy of the Carboneras area (see Fig. 3 for legend).

village El Saltador, also has tectonic slivers of marine late Burdigalian to Tortonian sediment (Figs 1A, 2A).

STRATIGRAPHY AND FACIES DESCRIPTION

Figure 2A gives the lithostratigraphic subdivision of the younger Neogene of the Nijar-Carboneras area. Messinian rocks were deposited in one major sedimentary cycle. The subdivision in a lower Turre Formation, a middle Yesares Formation, and an upper Feos Formation is primarily based on richness in marine fossils, (bio)chemical precipitates and terrigenous debris.

Figure 2B gives the relation of my lithostratigraphic scheme to earlier work on part of the Northern Nijar Basin and the adjacent Vera and Sorbas Basins.

A. The Turre Formation

The Turre Formation is a complex of marine fossiliferous sediments that unconformably covers pre-Neogene basement along the northern and western basin margin and volcanic basement along its eastern margin (Figs 1 and 2A). Like Dronkert & Pagnier (1977), I recognize three members of the Turre Formation: the Azagador Sandstone at its base, grading upwards into the Abad Marl in more central parts of the basin and covered by the Cantera Reef Limestone along the basin margin (Figs 1, 2A, B).

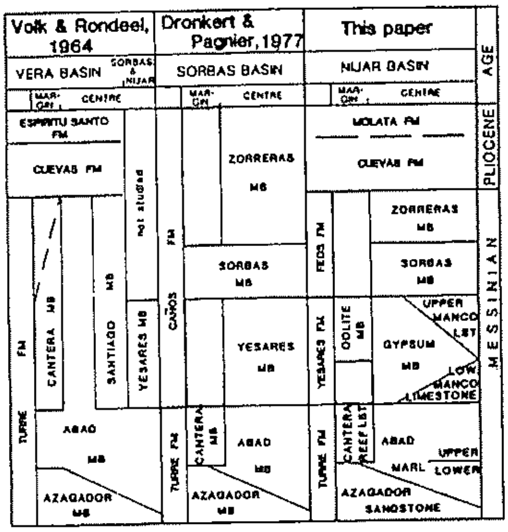


Fig.2B. Stratigraphic correlations with nearby basins.

The *Azagador Sandstone* consists of brown to yellow, mixed siliciclastic-bioclastic sandstone with basal conglomerates of coarse fragments of the directly underlying pre-Neogene or volcanic basement (Rondeel 1965, Van de Poel et al. 1984). Its top is especially rich in diverse marine macrofossils (mollusks, echinoids, Bryozoa, *Heterostegina* sp., calcareous algae and, locally, corals).

The *Abad Marl* is characterized by the dominance in foraminifera-rich marl, which, near the top of this unit, alternates with numerous soft, white, ca. 50 cm. thick diatomitic beds (Fig. 3A; Plates Ia and IIa, b). Calcified or silicified diatomitic levels interrupt the marlstone at regular intervals in the lower part of this unit, while bioclastic calcarenite layers are present both near its base and in its upper part. The informally distinguished Upper and Lower Abad Marl contrast in the field by a darker, yellowish, versus a lighter, greyish colour, respectively. At a fresh surface, the boundary coincides with an upwards change from bluish-grey, massive, bioturbated marlstone to laminated yellowish pelitic sediment (marl, lime-mud, calcisiltite). The yellowish colour of the Upper Abad Marl is caused by a relatively high content in fine-grained iron-hydroxides, whereas, at the same time, its sediments are rich in microscopically small gypsum crystals (cf. Fe and SO_4 curves of Fig. 4). An upward increase in manganese-hydroxide gives rise to a number of conspicuous, dark-brown levels of a few centimeters thick and the Abad Marl terminates in a dolomitic interval (Fig. 3A; Mn and Mg-curves of Fig. 4).

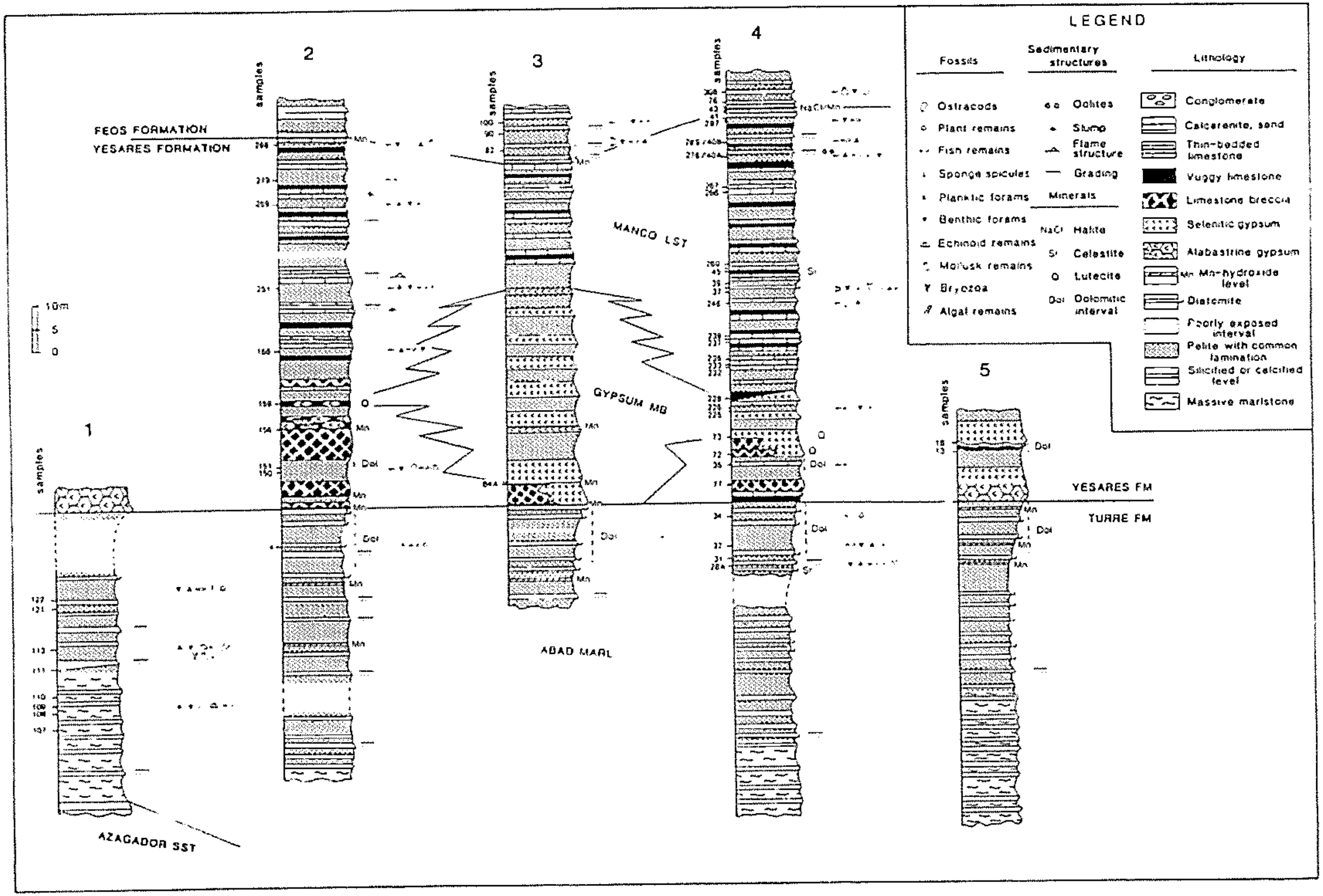


Fig. 3A. Lithologic sections of the Messinian Yesares Formation and associated deposits of the central part of the Northern Nijar Basin (see Fig. 1A for locations). Multiple intervals of 'gypsum-ghost limestone' (in black) characterize the Manco Limestone.

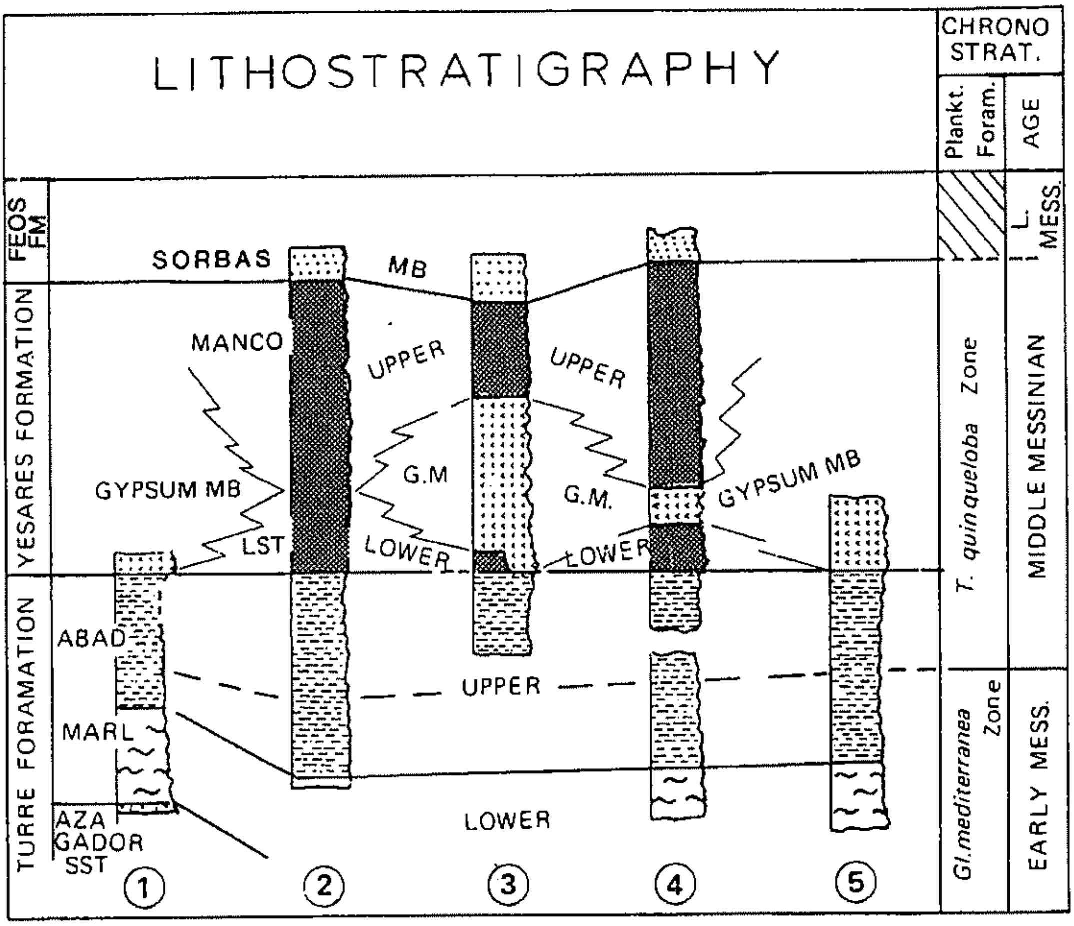


Fig. 3B. Lithostratigraphy, age and correlation of lithologic sections of Figure 3A, showing thinning of Abad Marl in westernmost section (1) and rapid lateral transitions between Gypsum Member and Manco Limestone.

Near the base of the Upper Abad Marl, the first of a number of bioclastic layers occurs, that are typical for this interval (Fig. 3A, Plate II b). Most of these beds rapidly wedge out and display normal grading, sometimes with parallel lamination and horizontal burrowing at the top. The richness in fragments of algae, Bryozoa, mollusks, and larger Foraminiferida attests to turbidity currents and debris flows originating from a carbonate shoal. Frequency and thickness of these layers increases in westerly direction (Fig. 3A). Besides, fine-grained turbidites, rich in shallow-water benthic foraminifers (*Ammonia beccarii* Linné, *Elphidium* sp., *Florilus boueanum* d'Orbigny, Miliolidae), ostracodes and fragile bryozoan remains, are common in the Upper Abad Marl. Such sediments dominate in the westernmost Upper Abad exposure, at El Puntal. Both in the latter locality and in the easternmost section (Los Vicentes, Fig. 1A), isolated slumped blocks of coral reef limestone have been found embedded in regularly laminated marl. The Abad Marl wedges out further to the west and north of El Puntal, where the Azagador Sandstone and Cantera Reef Limestone are directly superimposed (Figs 1, 2A, 9).

The Lower Abad Marl has a rich and diverse fossil content (Fig. 3A) and correlates with the lowermost part of the Messinian stratotype (Fig. 8). Foraminifer assemblages of the Upper Abad Marl are commonly dominated by small forms (Buliminacea, *Turborotalita quinqueloba* Natland). Remains of siliceous fossils (sponge spiculae, diatoms), and of plants and fish (Plate II c, d), increase in importance going upwards in the Abad section, to become virtually the only fossils present in the dolomitic top (Fig. 3A). A

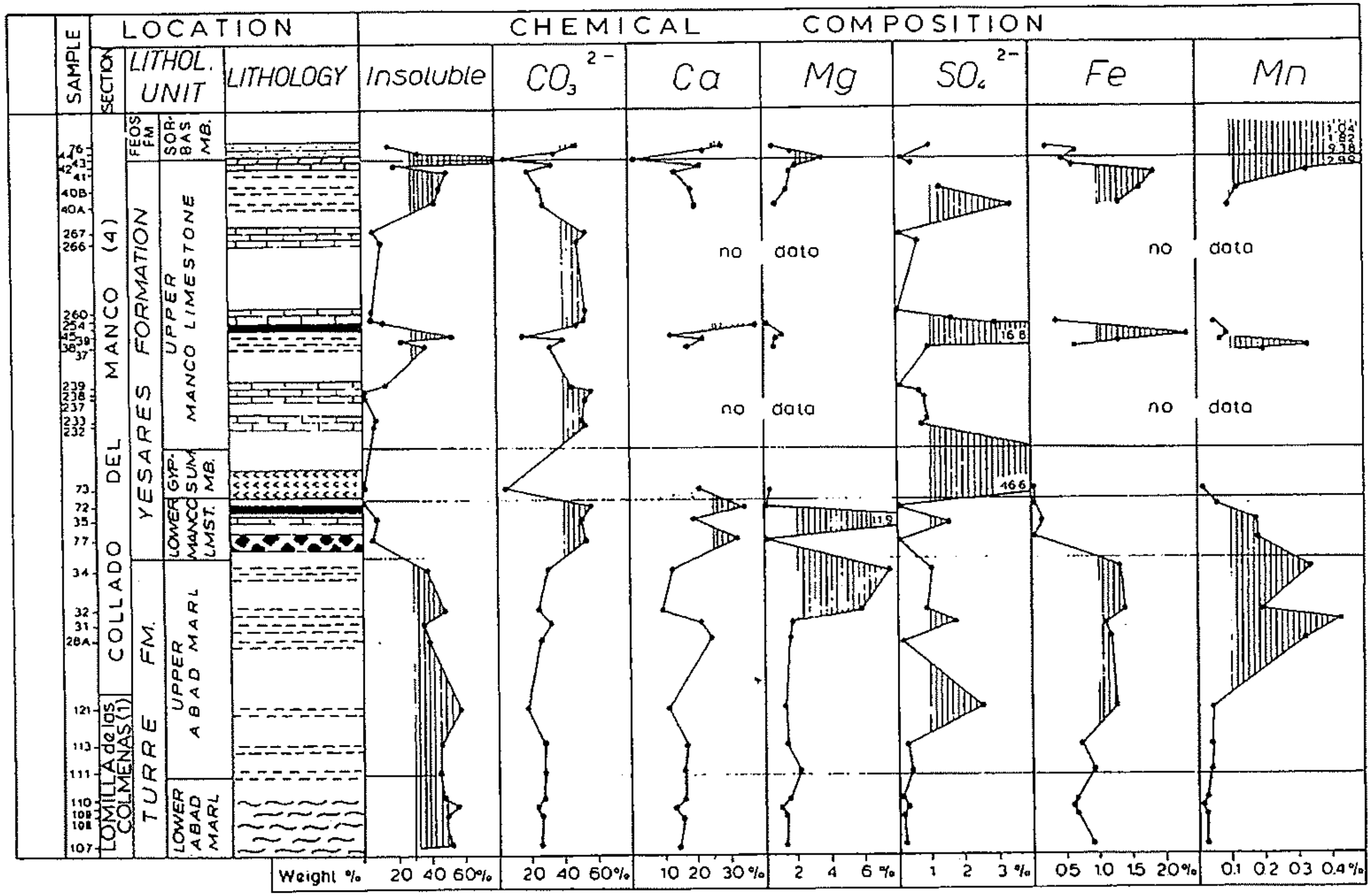


Fig. 4. Vertical distribution of major and some minor chemical constituents in the lower-middle Messinian of the central part of the Northern Nijar Basin (see Fig. 3 for legend). Note: (1) similar composition of Upper Abad Marl and laminated intervals of the Yesares Formation; (2) pure calcium carbonate composition of vuggy limestones and limestone breccia of the Manco Member.

consistent change in coiling direction of *Neogloboquadrina acostaensis* Blow, marking the boundary between the local planktonic foraminifer *G. mediterranea* and *T. quinqueloba* Zones (Fig. 3B), takes place some tens of metres below the top of the Abad Marl and is also found within the Messinian Tripoli Formation of Sicily (Fig. 8).

The *Cantera Reef Limestone*, formed by massive reef limestone with associated bioclastic calcarenite and calcirudite, only occurs in the northwestern part of the studied area. In sections most proximal to the basin margin, the Reef Limestone rests directly on Azagador Sandstone (e.g. Cerro de la Lancha and Molinillo) while, more basinward, it progrades over the Abad Marl, as can be seen in the vicinity of Polopos (Figs 1, 2A, 5, 9). Sediments in the same stratigraphic setting near Nijar (Fig. 1B) have been described as the 'Reef Complex' by Dabrio et al. (1981). They discerned three facies that can also be recognized in our area (Fig. 5): 1) massive reef core with dominance of the coral *Porites* sp. (extensively exposed at the top of the hills along the higher streambed of the Rambla Lucainena); 2) reef fore-slope deposits with high-angle giant cross-bedding (well-exposed in the southern bank of the Barranco del Pino); c) light-coloured, bedded calcarenite and calcirudite, representing the toe of a slope-facies (hills around Polopos).

Due to erosion, that already started in the late Messinian (Fig. 9C), the continuous transition from reef to basinal facies is not preserved. As yet, the Cantera Reef Limestone can be correlated with the Upper Abad Marl since: (a) the west to east distribution of reef to slope subfacies of the Cantera Reef Limestone can be matched with an east-west increase in importance of reef debris in the Abad Marl; (b) Cantera deposits prograde around Polopos over older Abad Marl, which also underlies Upper Abad Marl in more central parts of the basin; (c) the coral faunas of the Reef Limestone and of the Upper Abad Marl are both oligotypic; (d) in the adjacent Agua Amarga Basin, both Upper Abad Marl and Cantera Reef foreslope deposits are conformably overlain by a coarse breccia of limestone with evaporite-relics (Van de Poel et al. 1984).

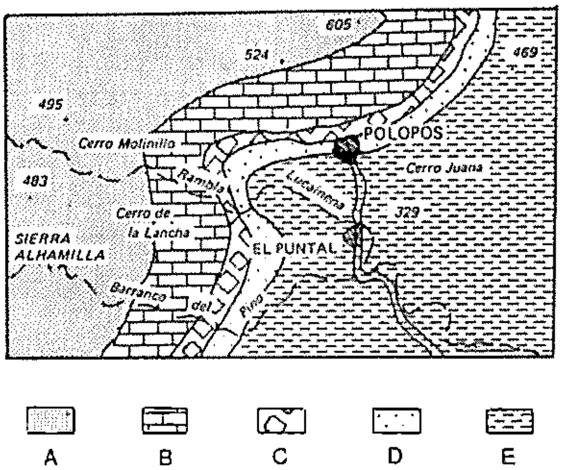


Fig. 5. Distribution of youngest Cantera Reef Limestone subfacies in the Polópos area (partially reconstructed). A = older formations; B = massive reef limestone; C = proximal fore-reef; D = distal toe-of-slope facies; E = basinal Upper Abad Marl. Numbers refer to hilltop elevations in meters.

B. The Yesares Formation

The Yesares Formation is characterized by thick-beds of evaporite or of breccious or massive limestone with 'ghosts' of gypsum (Figs 3A, 6, 7; Plates I, IIa, III). In the central part of the basin, the Yesares Formation covers the Abad Marl conformably and is intercalated with laminated pelitic sediments that are lithologically similar to the Upper Abad Marl (Figs. 3A, 4). At the NW basin margin, a series of oolitic limestones and intercalated gypssiferous algal breccia, covers the Cantera Reef Limestone with an erosional contact (Fig. 6). The massive Messinian gypsum-bearing strata of southeastern Spain were originally referred to as Yesares Member of the Turre Formation (Völk & Rondeel 1964) or as Yesares Member of the Caños Formation (Ruegg 1964; Dronkert & Pagnier 1977) (Fig. 2B). Here, a separate formational status is given to the series that bears thick beds of evaporite or of limestone with ghosts of gypsum crystals. Most of the constituent facies are represented in the Loma de los Yesares, the type-area designated by Völk and Rondeel (1964) for their Yesares Member (cf. Section 3 of Figs 1 and 3). The formation is especially well exposed in Sections 2 and 4 at the western and eastern end of the type-area, respectively (Plate I a, b, IIa).

The Yesares Formation is subdivided in three Members: the Oolite Member of the NW basin margin (Figs 1 and 6), and the Gypsum Member and Manco Limestone of the central part of the basin (Figs 1-3).

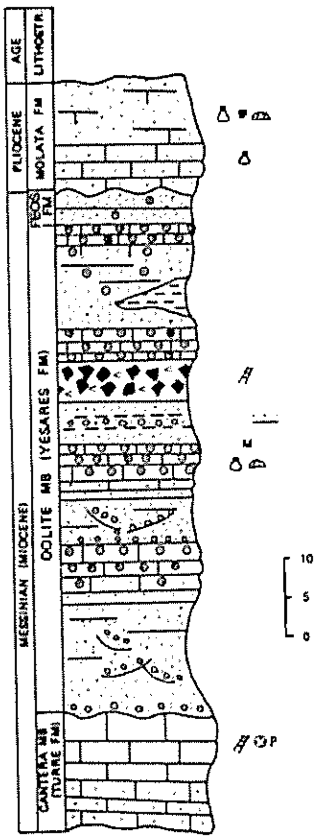


Fig. 6. Type section of the Oolite Member in the Barranco del Pino. P = *Porites* coral; M = *Microcodium*; For further legend, see Figure 3. For location see Figure 1 (section 6).

The *Oolite Member* is a mixed clastic-evaporitic series with oolitic limestone as its most characteristic element (Fig. 6). It is only found onlapping Cantera Reef Limestone along the northwestern basin margin (Figs 1, 2A, 9). The oolite-rich facies is especially well developed about 1 kilometer SW of the hamlet El Puntal (Fig. 1). A section in the northern bank of the Barranco del Pino serves as type-section (Section 6 of Figs 1 and 6). Here, the karstic top of the Cantera Reef Limestone is covered by a decimeters thick basal conglomerate, containing up to 6 cm. large pebbles of the pre-Neogene basement. In the overlying 50 meters or so, carbonatic intervals, consisting of oolitic limestones that may display mega-crossbedding, and a thick bed of grey-yellow limestone breccia. alternate with terrigenous clastics The breccia consists of stromatolite fragments of up to one decimeter, containing vugs that can have the form of gypsum crystals, and in which gypsum sometimes is preserved.

Some limestones of the Oolite Member contain abundant mollusk and echinoid remains. *Microcodium* is occasionally present at the top of the carbonate intervals. The terrigenous-clastic intercalations are barren of fossils except for scattered reworked fragments. These clastic intervals largely consist of material derived from the pre-Neogene metamorphic basement as exposed on the nearby slopes of the Sierra Alhamilla. Brown-red to purplish colours are common in the graded and mega-crossbedded fine conglomerates and sands and laminated silts.

Similar oolitic series have been described as 'Terminal Carbonate Complex' by Dabrio et al. (1981) from localities near Nijar and by Esteban & Giner (1980) from margins of the adjacent Agua Amarga Basin (Fig. 3.1B). The transition from Oolite Member to basinal facies is not exposed (Figs. 1, 9). As yet, the Oolite Member is considered the basin margin facies of the Yesares Formation, because of its stratigraphic position on top of the Turre Formation and its alternation of evaporitic, restricted marine and non-marine sediments (Fig. 2A; Esteban & Giner 1980, Dabrio et al. 1981).

The *Gypsum Member* is characterized by thick beds of massive gypsum (Figs 3A, 7; Plates Ia, b, IIIa, b). In good exposures these can be seen to be separated by important intervals of fine-grained, laminated pelite (e.g. base of Section 5, Plate IIIb). The middle part of Section 3 is the type-section for the Gypsum Member. It contains 8 gypsum layers with an upward decrease in bed-thickness. The Gypsum Member is here overlain by three cycles of Manco Limestone (Fig. 3A). In the SE flank of the Loma de los Yesares, 11 gypsum layers could be counted and no Manco Limestone is present (cf. Plate Ib, left foreground). The gypsum beds generally consist of coarse-crystalline translucent 'selenite' (Plate IIIb). Crystal-size decreases upwards within each individual layer. The base of the Gypsum Member commonly comprises fine-crystalline, white 'alabastrine' gypsum with chickenwire structure (e.g. Sections 1 and 5 of Fig. 3; Plate IIIa), which possibly contains relics of anhydrite (Dronkert 1985).

The Gypsum Member of the Northern Nijar Basin differs from that of the adjacent Sorbas Basin (Dronkert 1985) by the more common development of alabastrine subfacies, levels of manganese hydroxide and ferrous silica concretions. Small aggregates of lutecite (length-slow chalcedony) have been found in gypsum beds of the Collado del Manco (Section 4) and in rocks transitional from gypsum to limestone at Gafares (Section 3; Fig. 3A). Marly interbeds may contain levels of thin-bedded laminated limestone and minor amounts of vuggy limestone towards their dolomitic top (Plate IIIb).

The Gypsum Member is largely unfossiliferous but in some of the interbedded laminated pelites a restricted marine foraminifer microfauna (small-sized Buliminacea and planktonics) has been found, as in the interbeds of the laterally equivalent Manco Limestone (Figs. 3 and 7).

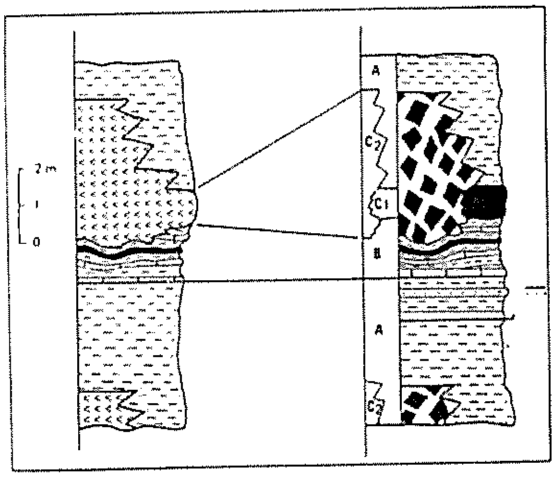
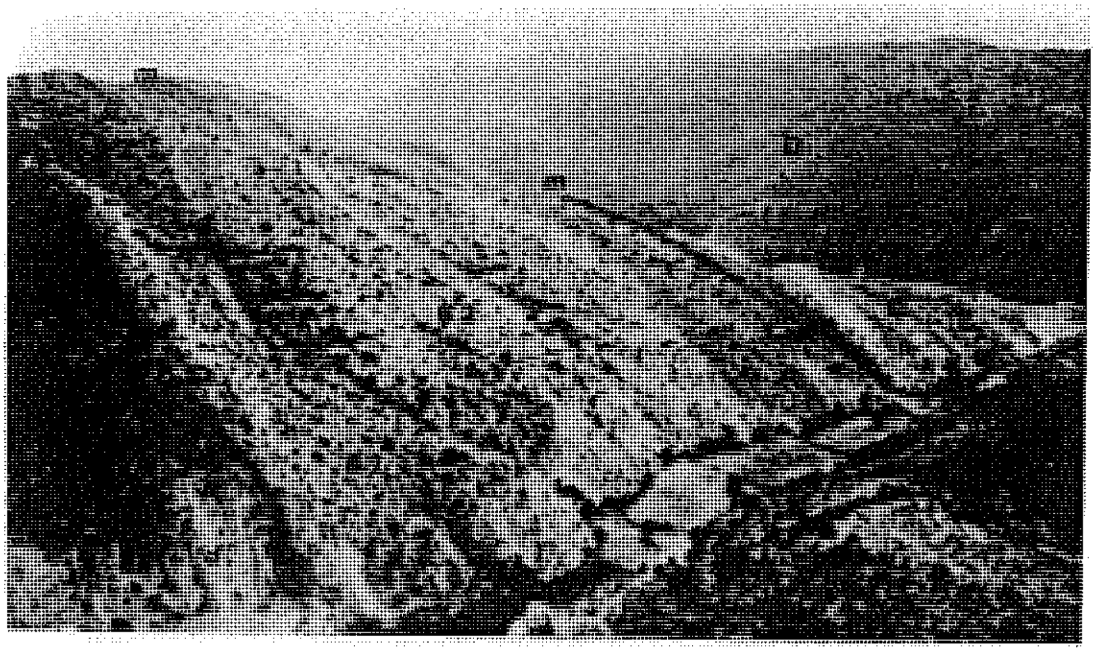


Fig. 7. Schematic sections to illustrate modal cycles in the Gypsum Member (left, slightly modified after Rouchy, 1981) and the Manco Limestone (right). Legend as in Fig. 3. C1 and C2 refer to the two gypsum-ghost limestone types. A more elaborate discussion on Manco Limestone subfacies is given in the text.

The *Manco Limestone* is characterized by thick beds of unfossiliferous, vuggy limestone or limestone breccia (Figs 3A, 7; Plates Ia, b, IIa, IIId, f). Together, they are referred to as 'gypsum-ghost limestone', since sedimentary and mineralogical features attesting to the original presence of gypsum are common in both types of rock (e.g. Plates IIa, IIIc, d, f, IVb,

Plate I. Messinian stratigraphy of the central part of the Northern Nijar Basin with emphasis on the Manco Limestone: **a.** Collado del Manco Section (4 of Figs. 1 and 3), with type-section for the Manco Limestone. From lower left to upper right are exposed: (A) Upper Abad Marl with white diatomitic intercalations (d); (B1) Lower Manco Limestone, thick-bedded limestone breccia with white dolomitic intercalation; the lateral transition, towards the top of the hill, of the upper two limestone breccia beds, into thick-bedded gypsum (B2: Gypsum Member); (B3) Upper Manco Limestone, an alternation of medium to thin-bedded limestone levels (grey) with marl intercalations (light-coloured), marked at its top by the strongly manganiferous level (m); (C) Feos Formation; (D) Pliocene calcarenites (below) and limestone of the Cerro Molata; **b.** Panoramic view of the Collado del Manco area. Legend as in Fig. A except for: (a) alabastrine gypsum along the local base of the Gypsum Member. Note: the rapid lateral transition from thick-bedded gypsum (B2) (e.g. Section 5 in the left background and 3' in the left foreground) to gypsum-ghost limestone (B1 / B3) (Sections 4 and 4') (cf. Plate Ia and Figs 1, 3A, B). ————>



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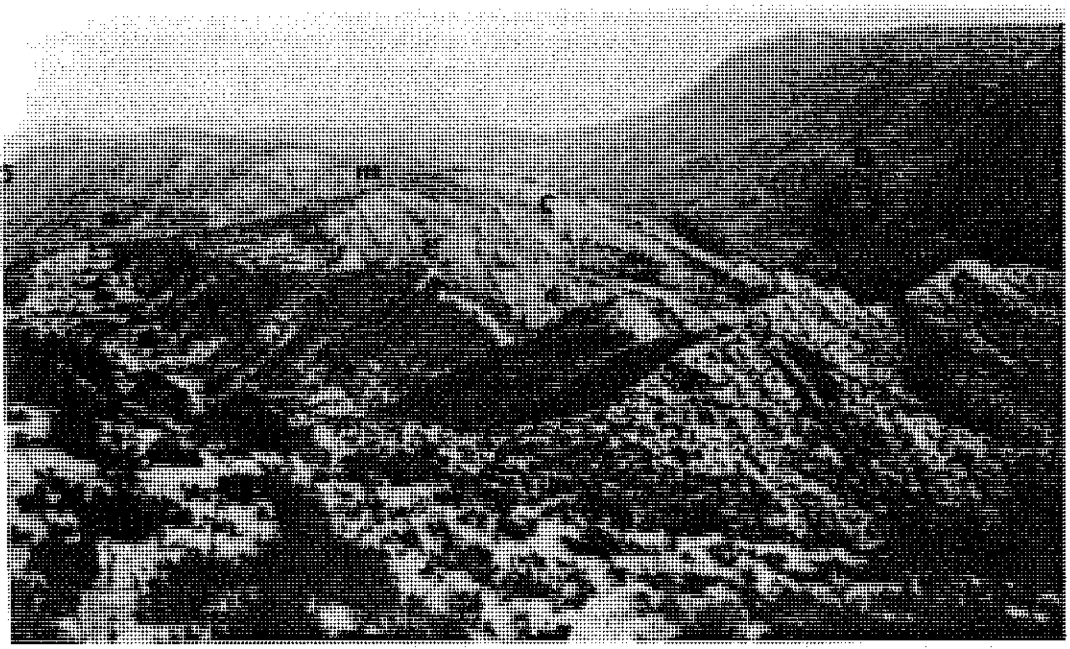


(b)

c, d, f). They are most common at the base or top of the Yesares Formation, but account for virtually the whole interval in the dry riverbed at Gafares (Section 2) and in the type section in the Collado del Manco (Pl. I; Section 4 of Figs. 1 and 3) (Van de Poel et al. 1984). The gypsum-ghost limestones alternate with meter-scale, laminated intervals (Fig. 3, Pl. IIId-f). Rouchy (1981) described cyclic sedimentation in the Gypsum Member from our area. A similar cyclic arrangement of sediments can be recognized within the Manco Member (Fig. 7). A medium to coarse-grained limestone breccia (*Facies C 2*) is the dominant lithology in the Lower Manco Limestone (Fig. 3; Plates I and IIa). Thick levels of thin-bedded, dolomitic, and medium-bedded, vuggy limestone (*Facies B and C 1*) dominate the Upper Manco Limestone (Fig.3; Pl. IIId-f).

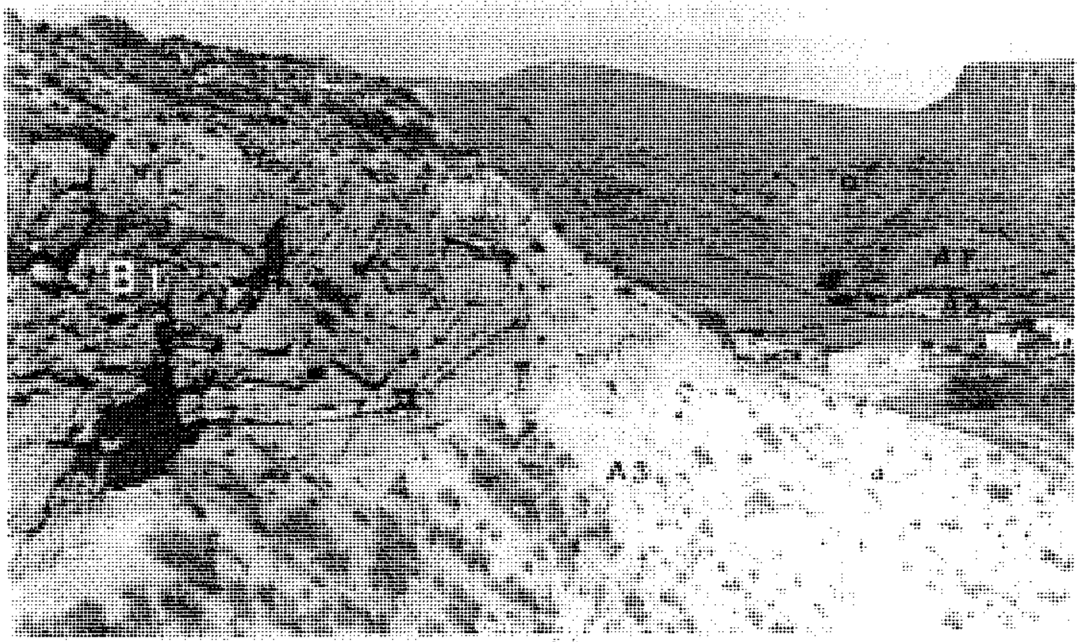
Facies A (Pl. IIId), the softer, marly lithology of the Manco Limestone is, like the Upper Abad Marl, laminated, displaying various colours, of which yellowish tints are the most common, and rich in dispersed microscopic gypsum crystals and iron and manganese hydroxides (Fig. 4). Oligotypical planktonic foraminifer assemblages consisting of small species (*T. quinqueloba* , *Neogloboquadrina* sp., *Globigerinita* sp.) occur and may be accompanied by small Buliminacea, and small Globorotaliidae (samples 166 C and 251 of Section 2). Shallow-water microfossils (*Elphidium* sp., smooth ostracodes, together with some miliolids and *Ammonia beccarii*) commonly are the dominant fauna constituents. Soft, white layers, rich in siliceous fossils (sponge spicules, diatoms) and bryozoan remains, occur in intervals 37-39 and 40A-B of Section 4 (Fig. 3A).

Facies B (Plates IIId-f, IVa, b), consists of white, soft dolomitic marl, commonly containing numerous, hard, 4 to 10 cm. thick, dolomitic limestone beds. In the Upper Manco Limestone, this facies continues uninterrupted for several meters (Pl. IIIe). Commonly developed lamination in the dolomitic limestone (Pl. IVa), may be of wavy character, caused by the growth of small gypsum crystals in a still soft sediment (Plate I b) (Rouchy 1981). This rock-type is transitional to facies C 1. Rare siliceous fossils (sponge spiculae, silicified miliolids) are found.

Facies C 1 (Plates IIId, f, IVc-d) is a grey to yellowish, massive vuggy limestone or, less common, a fine-grained limestone breccia, which occurs in beds of some decimeters to one meter. In the massive type, numerous cavities occur, which commonly mould gypsum crystals (Pl. IVc, d). They are generally less than 1 centimeter in size and, within a single layer, may show an upwards decrease in size. Often, they are partially or completely filled with sparry calcite forming pseudomorphs after gypsum. Rocks consisting entirely of pseudomorphed gypsum (Pl. IVf) are rare. In the fine-grained limestone breccias (Pl. IVe), vugs are smaller and less common than in the massive limestone, and often of irregular form. A thin section from the base of the vuggy limestone layer of Pl. IVc, demonstrates that the fine-grained breccias formed by collapse of vuggy limestones after the disappearance of the gypsum crystals (Pl. IVd).

Some vuggy limestones contain abundant celestite, but geochemical analysis of these rocks mostly shows a pure calcitic composition with a Mg-content oscillating between 0.15 and 0.20 % (Fig. 4; pers. comm. J.A. McKenzie 1983). At the top of Section 4, a siliceous, laminated limestone layer contains cube-shaped moulds and crystals of halite. The same level has strong manganese hydroxide enrichment (up to over 25 %; Van de Poel & Klaver 1989). This Mn-hydroxide enrichment level (Fig. 3; Pl. Ia, b), marks the top of the Yesares Formation over a large area in the central part of the basin (from the Cortijada de los Feos

Plate II. Stratigraphy and facies of Lower Manco Limestone and Upper Abad Marl at Gafares (Section 2). Cm-scale for rock-samples. **a.** In the foreground: (B 1) Lower Manco limestone, thick limestone breccia bed with (at arrow) coneshaped downward projection at base (calcified gypsum bush); (A 3) Upper Abad Marl with numerous white diatomitic intercalations (d). In the background: (A 2): Base of Lower Abad Marl; (A 1) basal Azagador Sandstone of the Turre Formation and (0) pre-Neogene basement. **b.** Typical lithology of the Upper Abad Marl: laminated marl with (t) thick biocalcarenitic turbidite and (d) white diatomitic intercalations (Level 4). **c.** Large plant-fragment from laminated marl (level 122, Section 1). **d.** Small fish-skeleton from diatomite (level 4, Section 2). ------------>



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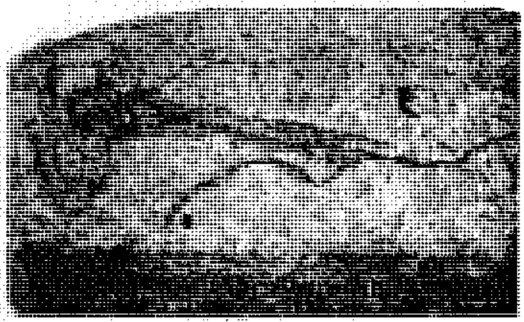


(d)

in the west up to the northern flank of the Cerro Molata in the east and the Rio Alias in the south, Fig. 3.1). Strong enrichment in manganese hydroxide is also common in Facies B and C 1 of the Lower Manco Limestone at Gafares (Fig. 3A; Pl. IVb, e). No fossils have been found in this facies.

Facies C 2 (Plates Ia, IIa, IVe) is a thick-bedded, medium to coarse-grained limestone breccia, composed of unsorted angular to subrounded fragments, which commonly show a saccharoidal microtexture. Components attain decimeter-size and lutecite aggregates, or small cavities moulding them, occur. Mineralogical composition is otherwise the same as in Facies C 1. No fossils have been found in this facies.

Additional facies. All Manco Limestone rocks of Section 2 (Arroyo Gafares) are relatively rich in fine siliciclastic detritus. Sedimentary structures as lateral thinning (Pl. IIId), distorted and graded bedding, with occasional flame structures and mudclasts along the base and parallel or cross-laminations at the top of the beds are here common (Fig. 3A).

All Manco Limestone sections show an increase in fine terrigenous debris towards their top, where graded calcarenites, embedded in soft, white laminites with some small *Elphidium* sp. and *Cyprideis* sp., are found. This lithology heralds the overlying Feos Formation.

The Manco Limestone correlates with the Messinian 'Calcare di Base' of Sicily (Fig. 8) for the following reasons: 1) it overlies, in stratigraphic continuity, diatomaceous Upper Abad Marl, which is correlative with fossils and lithofacies with the Sicilian Tripoli Formation; 2) similarly to the Calcare di Base it forms a lateral equivalent of 'middle' Messinian evaporites and is overlain by gypsum arenites, followed by a series with the typical late Messinian 'lago mare' fauna. Moreover, the Manco Limestone displays the same primary facies types and has the presence of celestite in common with the Calcare di Base (Richter Bernburg 1973, Decima et al. 1988).

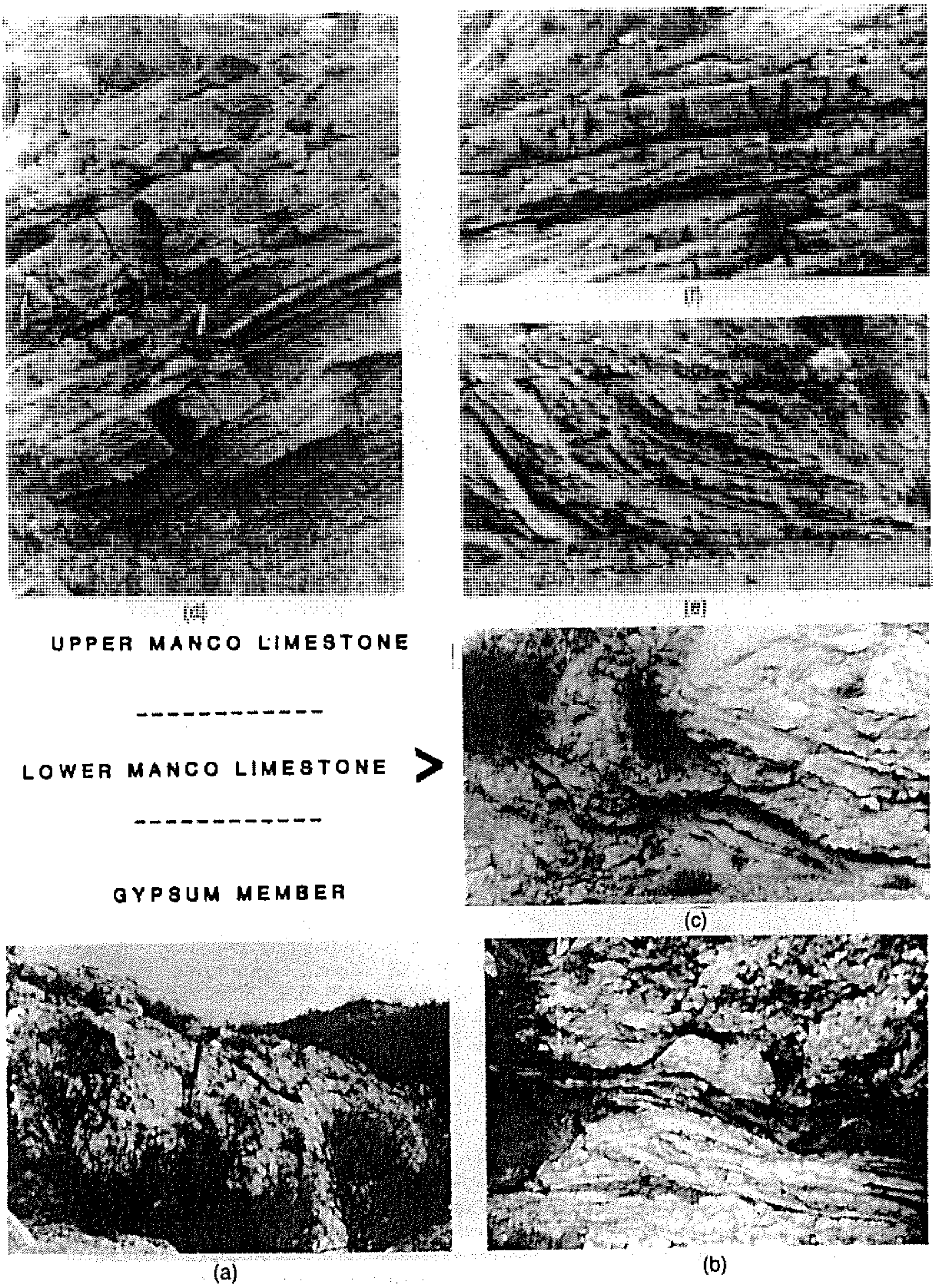
The Manco Limestone differs from the 'Calcare' in the the rarity of halite voids and the absence of elemental sulfur and unambiguous algal remains. In addition, a uniformly negative oxygen isotope signal (Table 1), as well as common manganese hydroxide-enrichment and interbeds of laminated marl with marine microfossils seem characteristic of the Manco Limestone.

C. The Feos Formation

In the central part of the basin, the Yesares Formation is conformably overlain by white, commonly laminated, fine-grained sediments with numerous coarser, clastic intercalations: the Feos Formation (Feos Member of Van de Poel et al. 1984) (Figs 1-3, 9; Pl. I a, b). It is subdivided in the Sorbas Member (below) and the Zorreras Member (above):

Sorbas Member. The lower part of the Feos Formation consists of laminated pelite with calcarenitic intervals containing ochreous, siliceous, thin laminated limestone beds (Fig. 3A). This interval is comparable in facies and stratigraphic position with the Sorbas

Plate III. Subfacies of Gypsum Member and Manco Limestone (hammer for scale): **a.** Alabastrine gypsum with chickenwire structure (base of Gypsum Member of Section 1); **b.** Thin-bedded alternations of pale dolomitic marl and limestone, overlain by selenitic gypsum with downward projecting cone (level 13-18, Section 5); **c.** Calcitized gypsum bed with cones along lower contact (Section 2, level 159); **d.** Typical subfacies of the Upper Manco Limestone (from bottom to top): (A) Laminated marl with turbidite intercalations (below hammer); (B) thin-bedded to laminated dolomitic limestone; (C) medium-bedded vuggy limestone (above hammer) (Section 2). **e.** Several meters of thin-bedded laminated dolomitic limestone (Facies B). Note slumping. (Upper Manco Limestone, Section 2); **f.** 30-cm thick bed of vuggy limestone (facies C 1) within thin-bedded dolomitic limestone (facies B). (Upper Manco Limetone, Section 2). ————>


(f)
(d)
(e)
UPPER MANCO LIMESTONE
LOWER MANCO LIMESTONE
GYPSUM MEMBER
(c)
(a)
(b)

Member of Ruegg (1964, in Dronkert & Pagnier 1977) from the Sorbas Basin (Fig. 2B).

The characteristic interbedded calcarenites consist of carbonate intraclasts, fragments of laminated and vuggy Manco Limestone, oolites and fragments of stromatolitic limestone. Gypsum arenites are common near the base of this unit. Slump structures and graded bedding, sometimes with flutecasts at the base of the beds, attest to the importance of gravitative sedimentation (Roep & Van Harten 1979).

Some of the soft, white laminated marls contain a few small *Elphidium* sp. and *Cyprideis* sp.. The coarser beds contain scattered reworked fossils. In sample 208 of Section 3, a rich 'lago mare' fauna (ostracodes and small mollusks) was identified.

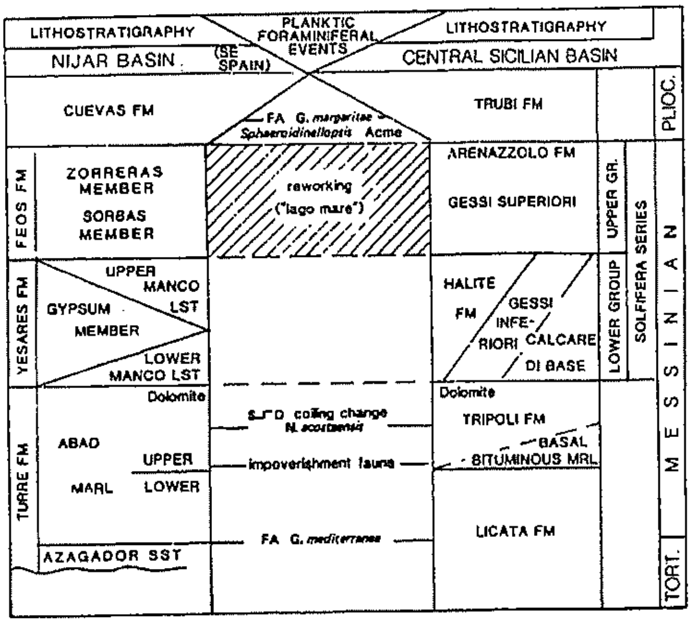
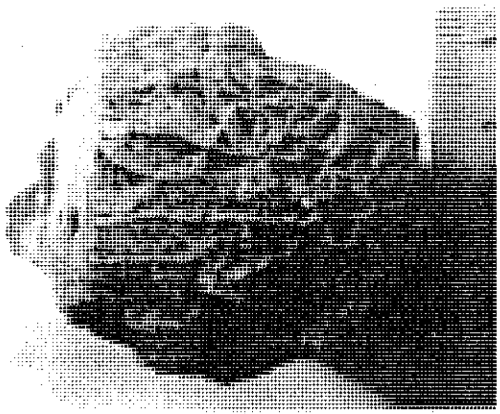


Fig. 8. Suggested litho-, and biostratigraphic correlation between the Messinian of the Nijar Basin and the Messinian of the stratotype area in Sicily (basal contacts of respective evaporite and "lago mare" units may not be exactly synchronous). Data on Sicilian stratigraphy from: Decima & Wezel (1973), Colalongo et al. (1979), McKenzie et al. (1979) and personal observations. No vertical scale implied.

Zorreras Member. The upper part of the Feos Formation is referred to as Zorreras Member, since it is equivalent in facies and stratigraphic position to the main part of the unit described under this name from the Sorbas Basin (Ruegg 1964, in Dronkert & Pagnier 1977; Fig. 2B). Three subfacies can be discerned:

1) Fluvial fan deposits, rich in pre-Neogene basement debris and fragments of Messinian reef and oolitic limestone, associated with thick, reddish, knobbly levels of caliche soil formation and relatively thin, but widespread, white laminited intervals. This subfacies is abundant west of Gafares (Roep & van Harten 1979, Dronkert et al. 1979) and has a strongly erosive base to the west and north of Los Feos, whereas it unconformably covers early Messinian Abad marls at El Puntal, near the western basin margin.

Plate IV. Sedimentary structures and microfacies of the Manco Limestone (cm-scales for rocksamples): **a.** Laminated dolomitic limestone (Facies B) with cellular structure (level 235, Section 4); **b.** Thin section of laminated limestone with small gypsum crystals (as voids up to 2 mm), which distort laminations into wavy forms. Black due to manganese hydroxide (transitional between facies B and C1, Section 2, level 156); **c.** Heavily vugged limestone (Facies C 1) with fine-grained breccia in base (Section 4, level 228); **d.** Thin section from bottom of sample of Plate IV c, showing brecciated vuggy limestone in the base. Largest diameter of vugs (light) is 7 mm; **e.** Limestone breccia (Facies C 2) with Mn-hydroxide coatings. (Section 3, level 84 A); **f.** Coalescent large pseudomorphs of calcite after gypsum (Section 2, level 159). ———>



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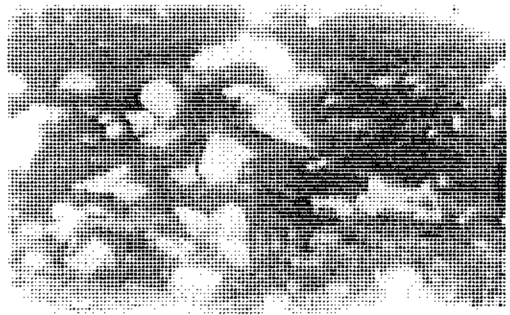

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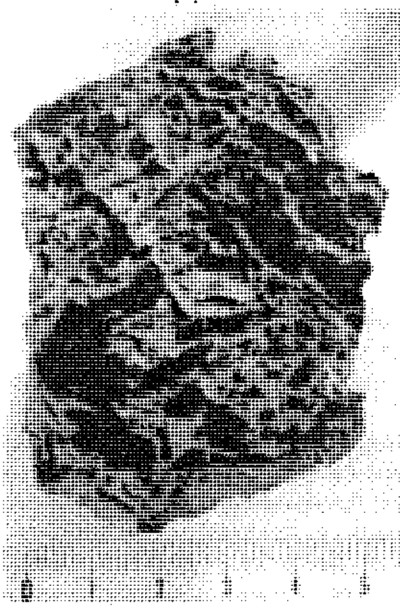

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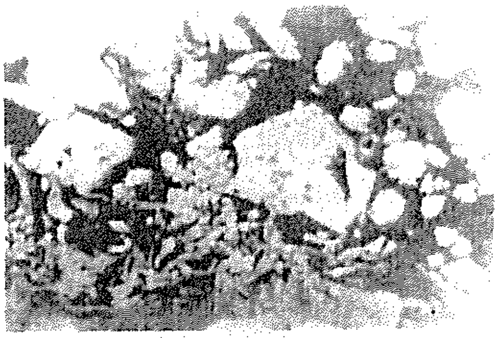

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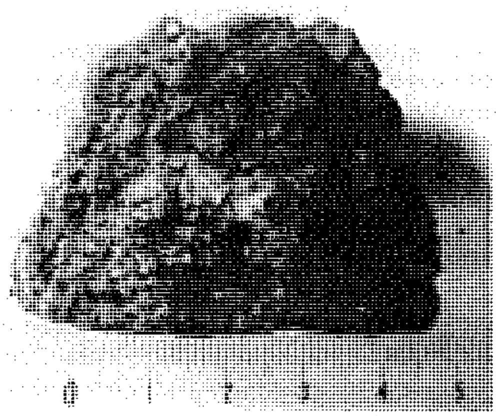

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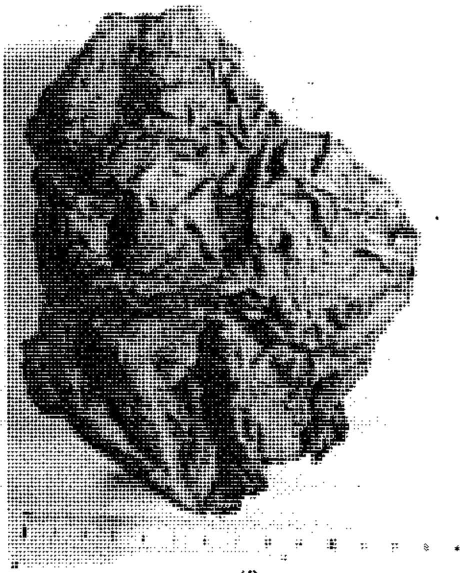


(f)

2) Green conglomerates, rich in well-rounded pebbles from the volcanic basement exposed to the east of the basin (Figs 1, 9C), alternating with white, laminated pelite. This subfacies is common in, and restricted to, the eastern part of the basin (sections around the Cerro Molata and Cerro del Caballon).

3) Light-coloured marly sediments with intervals of thin-bedded, laminated limestone and characteristic, white, carbonate intraclast conglomerates, that contain relatively few pre-Neogene or volcanic debris. This inter-fan subfacies is found between Gafares and the Cerro Molata.

Fine-grained sediments of the Feos Formation often contain ostracodes (*Cyprideis*) and the foraminifers *Ammonia beccarii* and *Elphidium* sp.. In the uppermost levels, these are associated with the ostracode genera *Loxoconcha* and *Tyrrhenocythere* (det. D. van Harten), small lamellibranchs, gastropods and the oogonia of *Chara* sp. This brackish to fresh biofacies is characteristic of the late Messinian 'lago mare' (e.g. Ruggieri & Sprovieri 1976, Colalongo et al. 1978). Its late Messinian age is confirmed by a local, conformable cover with marine silt of the Cuevas Formation (Figs 2A, B), containing a *Sphaeroidinellopsis-Globorotalia margaritae* planktonic foraminifer fauna of earliest Pliocene age (Fig.8). The marls of the Feos Formation also contain marine planktonic and benthic foraminifers reworked from the older Messinian.

DISCUSSION

Evolution of the depositional environment

Early Messinian.

The upward change within the Abad Marl from massive bioturbated marlstone with a diversified microfauna to laminites with oligotypical fossil assemblages attests to the initiation of restricted marine conditions in the earlier part of the Messinian (Troelstra et al. 1980).

Several phenomena suggest that oxygen-deficiency was the main feature of this restricted marine environment: (1) laminations, which can only be preserved in waters with an oxygen content about one tenth of normal and (2) dwarfism and diversity reduction of foraminifera, with strong dominance of the Buliminacea in the benthic assemblages (e.g. Harman 1964); (3) abundant fine plant material, well-preserved large plant fragments and entire fish skeletons (Pl. IIc, d); (4) authigenic microscopic gypsum crystals, common in organic-rich environments (Robert & Chamley 1974, Siesser & Rogers 1976, Briskin & Schreiber 1978).

The persistence of planktonic foraminifers during upper Abad deposition precludes a dramatic rise in salinity because these organisms have limited maximum tolerance values (ca. 50 ppt: Reiss & Hottinger 1984). The dominant *Porites* in the Cantera Reef Limestone indicates similar maximum salinity values (Kinsman 1964).

Middle Messinian (Fig. 9A, B).

Important hypersaline periods are recorded by the massive gypsum beds and the gypsum-ghost limestones of the Yesares Formation (Figs 3, 7, 9).

Interbedded marly sediments (Facies A of the Manco Limestone), with similar litho- and bio-facies as the Upper Abad Marl (Figs 3A, 4), represent nearly normal saline, but oxygen-deficient interruptions. At times, open-marine conditions may have been approximated, because some planktonic foraminifer faunas include small globorotaliids. Levels with abundant echinoid and mollusk remains from the marginal oolitic limestone similarly record important marine ingressions (Fig. 9B).

Periodic emergence and erosion of the marginal oolitic series, on the other hand, can probably be correlated with gypsum deposition in the basin centre (Esteban & Giner 1980, Dabrio et al. 1981). Emergence indicated by *Microcodium* at the brecciated top of oolitic

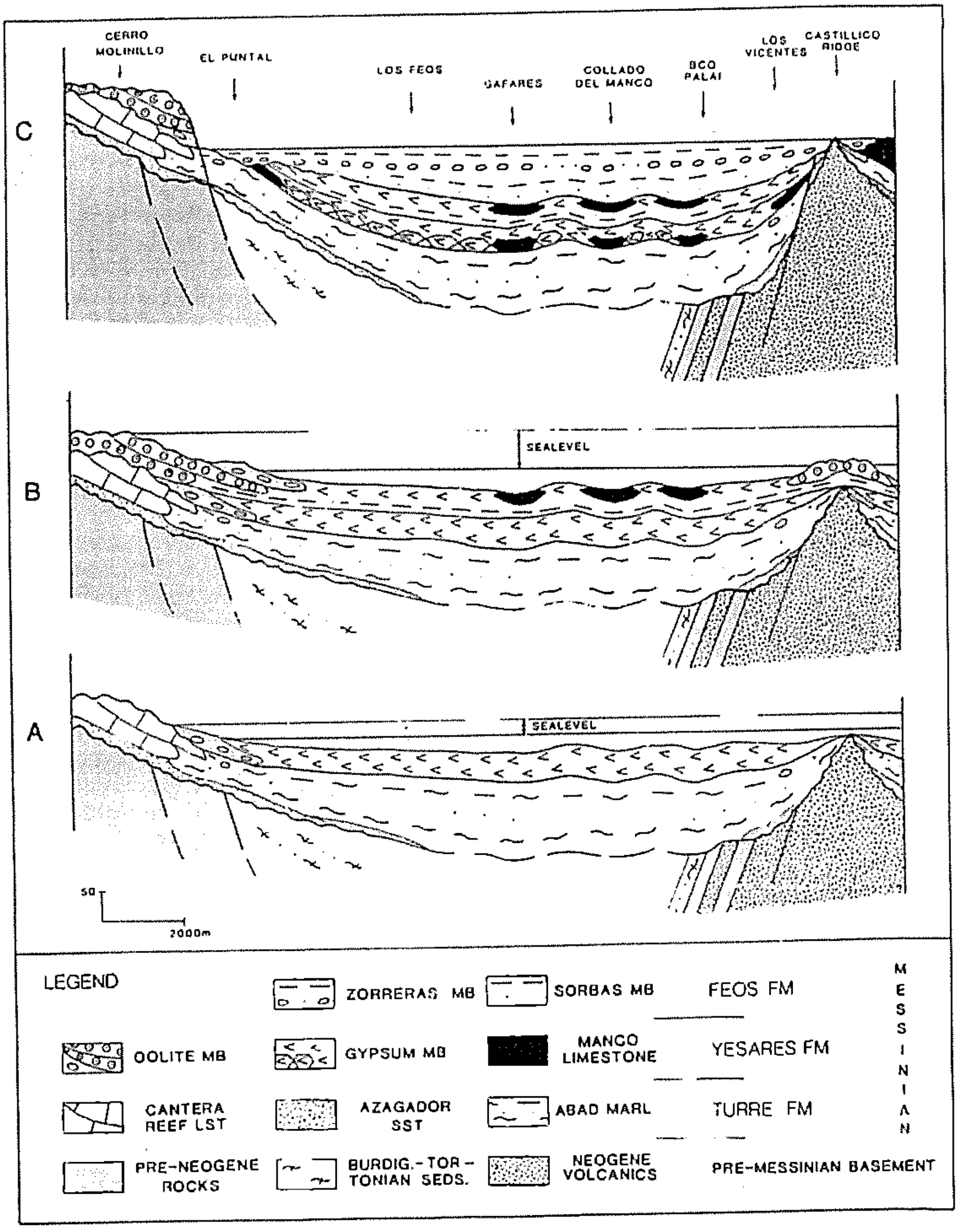


Fig. 9. Proposed model for the evolution of the Messinian depositional environment of the Northern Nijar Basin with emphasis on the Manco Limestone. 'Stage A': early Messinian oxygen-deficient conditions result in the deposition of laminated marly sediments in the basin centre; 'Stage B': middle Messinian alternation of evaporitic and oxygen-deficient conditions results in deposition of gypsum (?+anhydrite), gypsiferous and laminated limestones and laminated marly sediments in the basin centre; 'Stage C': late Messinian brackish 'lago mare' in the very basin centre and subaerial exposure in its more external parts results in dissolution of gypsum crystals and alteration of thick-bedded sulphate rocks from the underlying Yesares Formation to limestone breccia and (?)alabaster gypsum.

intervals in the Barranco del Pino area (Section 6) was the precursor to local fluviatile-deltaic deposition recorded by thick siliciclastic intercalations.

Late Messinian.
The composition and sparseness of the fauna in the basal part of the Feos Formation indicate hypersaline or brackish conditions. Hypersalinity may well have been maintained through the reworking of Messinian evaporites (gypsum turbidites).

The composition of the clastic intercalations and the existence of hiatuses attest to repeated subaerial exposure and the dominance of erosion in more external parts of the basin centre (Fig. 9C). The considerable fresh-water supply to the most central part of the basin, indicated by common brackish 'lago mare' faunas and intercalation of river-laid conglomerates along its margin reflect increasingly humid climatic conditions towards the end of the Feos deposition.

Origin of the gypsum-ghost limestones

The unfossiliferous nature, the narrow stratigraphic relation to massive gypsum beds, as much as the features they have in common with the latter which attest to the former presence of gypsum, all indicate that the gypsum ghost limestones were originally deposited in a strongly evaporitic environment. The lateral transitions between these limestones and the gypsum can be explained by assuming either original 'sedimentary' facies transitions or later, localized alteration of original gypsum beds. Since both, sedimentary facies transition and late alteration of gypsum, can have two different modes, four 'mechanisms' can hypothetically explain the common rapid lateral substitution of Messinian massive gypsum beds by gypsum ghost-limestone in the Carboneras area (Table 2). Their composition of low-magnesium calcite, which is not the normal $CaCO_3$ phase in a marine evaporitic environment (Friedman 1972, Decima et al. 1988), indicates that the original sediment underwent some type of diagenesis.

'Sedimentary' formation
Data on the original sedimentary environment and on the respective distribution of limestone and gypsum are used to evaluate wether environmental conditions were originally present to promote an early formation of gypsum-ghost limestone and, if so, in attempting to discriminate between the two 'sedimentary' mechanisms (A1 and A2 of Table 2). The respective distribution of Manco Limestone and Gypsum Member within the Yesares Formation is rather irregular (Figs 1, 3, 9C). Facies distribution within the Yesare Formation and in the under- and overlying rock units suggests that, the Upper Manco Limestone, in particular, represents a relatively deeper sedimentary environment than the Gypsum: (1) it is concentrated in the eastern-central part of the basin, where the underlying Abad Marl reaches its greatest thickness and predominantly consists of fine-grained sediment (Figs. 1-3, 9) and (2), it is overlain by a thick, predominantly fine-grained, Feos Formation, while the Gypsum Member in the western-central area is covered by a Feos series in which continental red-bed and fluviatile facies dominate (Fig. 9 C). There are no erosional surfaces and no unambiguous shallow-water indicators

Table 1. Isotopic data for Lower Manco Limestones*

Sample	Limestone lithology	$\Delta^{13}C$	$\Delta^{18}O$
1	massive	-5.12	-6.31
2	wavy laminated	-5.29	-6.65
3	breccia	-4.96	-7.06
4**	massive	-4.31	-5.46

*sample locality Arroyo Gafares (section 2)
**base of Lower Manco Limestone

Table 2. Different diagenetic pathways to produce low-magnesium calcite from an evaporitic sediment

Type	Name	Original material	Diagenetic agent	Final product
Early ('sedimentary')	A1 (Neev & Emery 1967, Friedman 1972)	'basinal' gypsum	sulfate-reducing bacteria + organic C	calcite + H_2S
	A2 (Decima et al.1988)	'marginal' aragonite	mateoric water ($+CO_2$)	calcite + strontianite
Late ('alteration')	B1 (Feely & Kulp 1957)	massive gypsum & anhydrite	sulfate-reducing bacteria + organic C	calcite + S_2
	B2 (West 1973)	massive gypsum & anhydrite	meteoric water + CO_2 ($+Sr^{2+}$)	calcite + celestite

in the Manco Limestone. Gypsum ghost limestones from its upper part are typically interbedded with marine marls and, locally, turbidites and slump facies (Arroyo Gafares).

The relative importance of reducing conditions during the formation of the original sediment of the vuggy limestone (Facies C1) becomes most apparent when its vertical facies relations are considered. Oxygen-deficiency is already indicated by the preservation of lamination, oligotypical microfaunas and an abundance of microscopic authigenic gypsum in the underlying Upper Abad Marl and in similar intercalated marls (facies A) of the Manco Limestone itself. The dolomitic intervals at the top of the Abad Marl and within the Manco Limestone (facies B), to which facies C1 is most closely associated (Figs 3A, 4, 7; Plates III d, f, IV b), probably reflect highly anoxic conditions, as early diagenetic formation of dolomite commonly is accompanied by bacterial sulphate reduction (Friedman 1966, Kelts & McKenzie 1982, Patterson & Kinsman 1982). McKenzie et al. (1979) and Bellanca & Neri (1986) provide stable isotope evidence of anoxic conditions during the formation of dolomitic intervals at the top of the Tripoli Formation and within the Calcare di Base from the Messinian of Sicily. These formations are equivalent in facies and stratigraphic position to the top of the Abad Marl and Manco Limestone, respectively (Fig. 8).

The stratigraphic setting of the gypsum ghost limestones, in particular those from the Upper Manco Limestone, thus is highly suggestive of an original formation in a relatively deeper, anoxic part of the basin as prerequisite for 'mechanism A1', and early bacterial sulfate reduction probably was a first instrument in producing limestone (Fig. 9B).

Later alteration

A late bacterial transformation of gypsum (mechanism B1) is considered less likely to have produced the gypsum ghost limestones of Carboneras, since they lack characteristic native sulphur and strong depletion in ΔC13 (Table 1; Dessau et al. 1962, Decima et al. 1988, Pierre & Rouchy 1988).

A later fresh-water diagenetic and dissolution stage for the vuggy limestones (facies C1) is indicated by the common voids moulding gypsum crystals, the occurrence of fine-grained dissolution-collapse breccias and the presence of celestite. Fresh water alteration of massive sulfate rock is considered to have been the essential mechanism (B2 of Table 2) to produce the thick-bedded coarse limestone breccias (facies C2) of the Lower Manco Member. Massive replacement of gypsum by limestone is documented by coalescing pseudomorphs of calcite after gypsum (Plate IVf) and in cone-shaped projections, at the base of thick Manco limestone beds, that are identical to features at the base of gypsum beds (Plates IIa, IIIb, c). Accessory lutecite is a further strong indicator for 'vanished evaporites' (Munier Chalmas 1890, Folk & Pitman 1971).

The distribution of coarse limestone breccia of the Manco Member is closely related to the nodular alabastrine gypsum subfacies of the Gypsum Member. Both are restricted to the base of the Yesares Formation and in a number of localities nodular alabastrine gypsum is found as intermediate in the rapid, lateral substitution of massive selenitic gypsum by thickly-bedded limestone breccia (e.g. eastern end of the Collado del Manco: Plate Ib). Alabastrine gypsum is commonly considered a product of fresh-water diagenesis of nodular anhydrite (e.g. Shearman 1966, Toulemont 1980).

The uniform, strongly negative oxygen-isotope values of limestones from the Lower Manco Member, lastly (Table 1), support a fresh-water diagenetic origin for the gypsum-ghost breccias.

Although paleontologic data do not exclude temporary brackish-water conditions in the basin centre during Yesares Formation deposition (Van de Poel, in prep.), a fresh-water diagenetic stage is most easily envisaged in the latest part of the Messinian, when there is ample evidence of abundant meteoric water in the centre of the basin and subaerial exposed Yesares beds in its somewhat more external parts (Fig. 9C). Major dissolution collapse of the entire Yesares Formation of the centre of the adjacent Agua Amarga Basin probably took place at this time (Van de Poel et al. 1984).

CONCLUSION

The formation of the gypsum ghost limestones of the Carboneras area was a multi-stage process. Most of the gypsum-ghost limestones originally formed as gypsum in the central part of the basin. During periods of low oxygen, sulphate-reducing bacteria rapidly transformed thinner gypsum beds, as especially developed in the upper part of the Yesares Formation into predominantly carbonatic sediment. In a later stage, substantial dissolution and alteration of gypsum took place when porous rocks of the Yesares Formation served as fresh-water reservoirs.

3.3. FORAMINIFERAL BIOSTRATIGRAPHY AND PALEOENVIRONMENTS OF THE MIOCENE-PLIOCENE NORTHERN NIJAR BASIN (S.E. SPAIN)[1]

H. M. van de Poel

SUMMARY

Five successive, distinct, microfossil assemblages, primarily characterized by their benthic foraminifer content, have been recognized in the Mio-Pliocene of the Northern Nijar Basin (S.E. Spain). The assemblages record a number of fundamental changes in the environment of the area, which have been dated by means of planktonic foraminifer biostratigraphy.

In the earliest Messinian the central part of the basin was several hundreds metres deep and had relatively open marine conditions, which were followed by oxygen-deficiency, accompanied by slightly deviating salinities and some reduction in waterdepth. Subsequent deposition of Messinian evaporites was interrupted by sparse more open marine conditions.

During the late Messinian brackish conditions became dominant and the basin filled with delta-plain sediments. The early Pliocene saw a sudden return to open marine, outer neritic conditions, which shoaled to an inner neritic environment in the middle Pliocene.

Since the Northern Nijar Basin was close to the connection between the Mediterranean Sea and the Atlantic Ocean, its history shows that Atlantic inflow became more and more severed in the course of the Messinian and was, either completely or virtually, obstructed in its latest part.

INTRODUCTION

Foraminifer biostratigraphy and paleoecology of the Mio-Pliocene of the southern Almería province (S.E. Spain; Fig. 1) have previously been discussed in detail for part of the section (Montenat et al. 1976, Civis et al. 1979, Troelstra et al. 1980, Poore & Stone 1981) or as part of studies with a wider framework (Völk 1967, Iaccarino et al. 1975, Perconig 1976, Geerlings & Van de Poel 1979, Dronkert et al. 1979, Ott d'Estevou 1980, De la Chapelle 1988, Van de Poel 1991).

Different opinions remain, in the first place as to the question whether Messinian evaporite deposition was interrupted and followed by important marine incursions in this area (Perconig 1976, Montenat et al. 1976, 1980, Ott d'Estevou 1980, Poore & Stone 1981, Müller & Hsü 1987, De la Chapelle 1988) or whether it remained isolated during middle and upper Messinian time (Iaccarino et al. 1975, Geerlings & Van de Poel 1979, Dronkert et al. 1979, Roep & Van Harten 1979, Troelstra et al. 1980, Geerlings et al.1980a, b, De Deckker et al. 1988).

[1]Text and figures earlier published in *Scripta Geologica 102 (1992), pp. 1-32.*

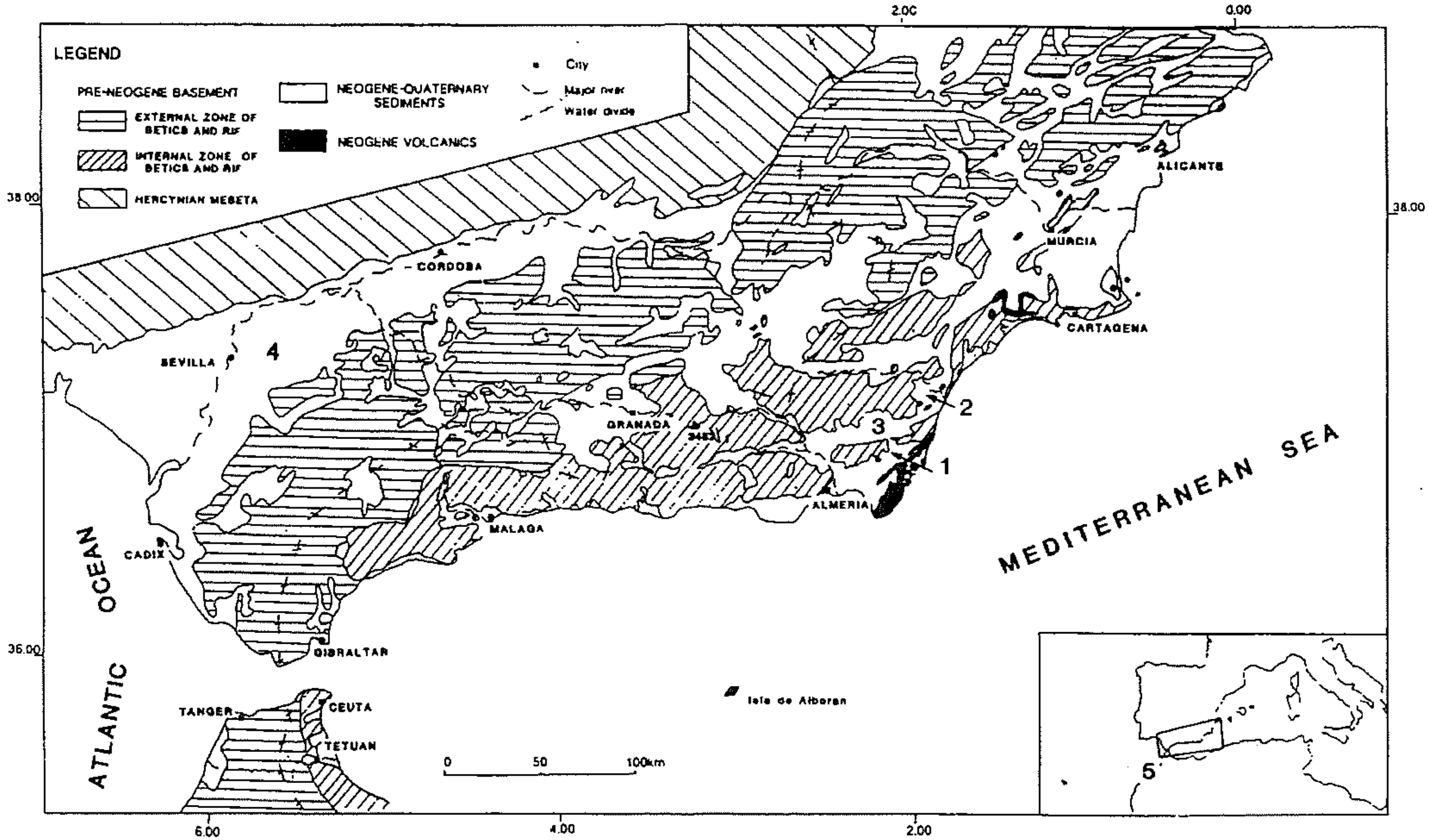


Fig. 1. Generalized geological map of the Betic Cordilleras (S. Spain) showing areas with Late Cenozoic sediments (in white), representing potential 'Mediterranean entrance areas'. Numbers refer to Neogene basins mentioned in text: 1 = Northern Nijar Basin; 2 = Vera Basin; 3 = Sorbas Basin; 4 = Guadalquivir Basin; 5 = Gharb Basin.

On a more regional scale, opinions converge on the importance of upwelling of nutrient-rich deeper water during deposition of the sediments directly underlying the Western Mediterranean Messinian evaporites (Bizon et al. 1979, McKenzie et al. 1980, Gersonde 1980, Van der Zwaan 1982, Müller & Schrader 1989, Hodell et al. 1989, Benson et al. 1991). No agreement has been reached on whether the Mediterranean shoaled during the early Messinian, nor on whether salinities were still normal or already elevated at this time (Sturani 1978, Cita et al. 1978, Colalongo et al. 1979, Orszag Sperber et al. 1980, Troelstra et al. 1980, Ott d'Estevou 1980, Poore & Stone 1981, Van der Zwaan 1982).

Since the Almeria region is situated at the Mediterranean end of the Atlantic/Mediterranean gateways (Fig. 1), its fossil record may provide answers to those questions (Müller & Hsü 1987, Benson et al. 1991).

This chapter describes the Mio-Pliocene microfaunal development, and discusses its paleoenvironmental and paleogeographical significance. The studied sections are located in the central part of the Northern Nijar Basin, which is a Neogene-Quaternary depression near Cabo de Gata (Figs 1 & 2). Details on lithology and stratigraphy of the sections were given in Chapter 3.2.

METHODS

A (composite) section in the eastern-central part of the Northern Nijar basin has been sampled in detail for micropaleontological analysis (Section 3 of Figs 2 & 3). Two other sections, one at Gafares (Section 2) and one near Los Feos (Section 1) were studied to check whether the observed trends are of more than local importance, and to get more information on the relation between microfossil content and lithology. The sampled interval (Fig. 3) comprises the Abad Marl of the Turre Formation, the Messinian evaporite-bearing Yesares Formation, the alternating coarse clastic and pelitic sediments of the Feos Formation, and the calcareous silts and sands of the Cuevas and Molata Formations, which are rich in marine macrofossil remains (Van de Poel, 1991).

Samples have been washed over a 63 μ sieve and have been examined to determine five semi-quantitative frequency classes of microfossil taxa and some non-microfossil constituents (Appendix IIa-c); details on the taxa are given in Appendix I. Ranges of characteristic species or genera are presented in Figure 4.

The relative frequency of major and some minor chemical constituents in a set of samples from the Abad Marl has been determined by atomic absorption spectrometry (Fig. 5).

BIOSTRATIGRAPHIC DESCRIPTIONS

ASSEMBLAGES

The distribution of microfossils (Appendix 1 and Fig. 4) allows for the recognition of 5 distinct Assemblages (A-E). The subdivision is primarily based on changes in benthic foraminifer content, but also takes the relative frequency and composition of planktonic foraminifer fauna and remaining fossil constituents, as well as the relative frequency of terrigenous detritus and some authigenic minerals into account. It is based on Section 3 (App. IIa), but can also be recognized in the other sections (App. IIb, c).

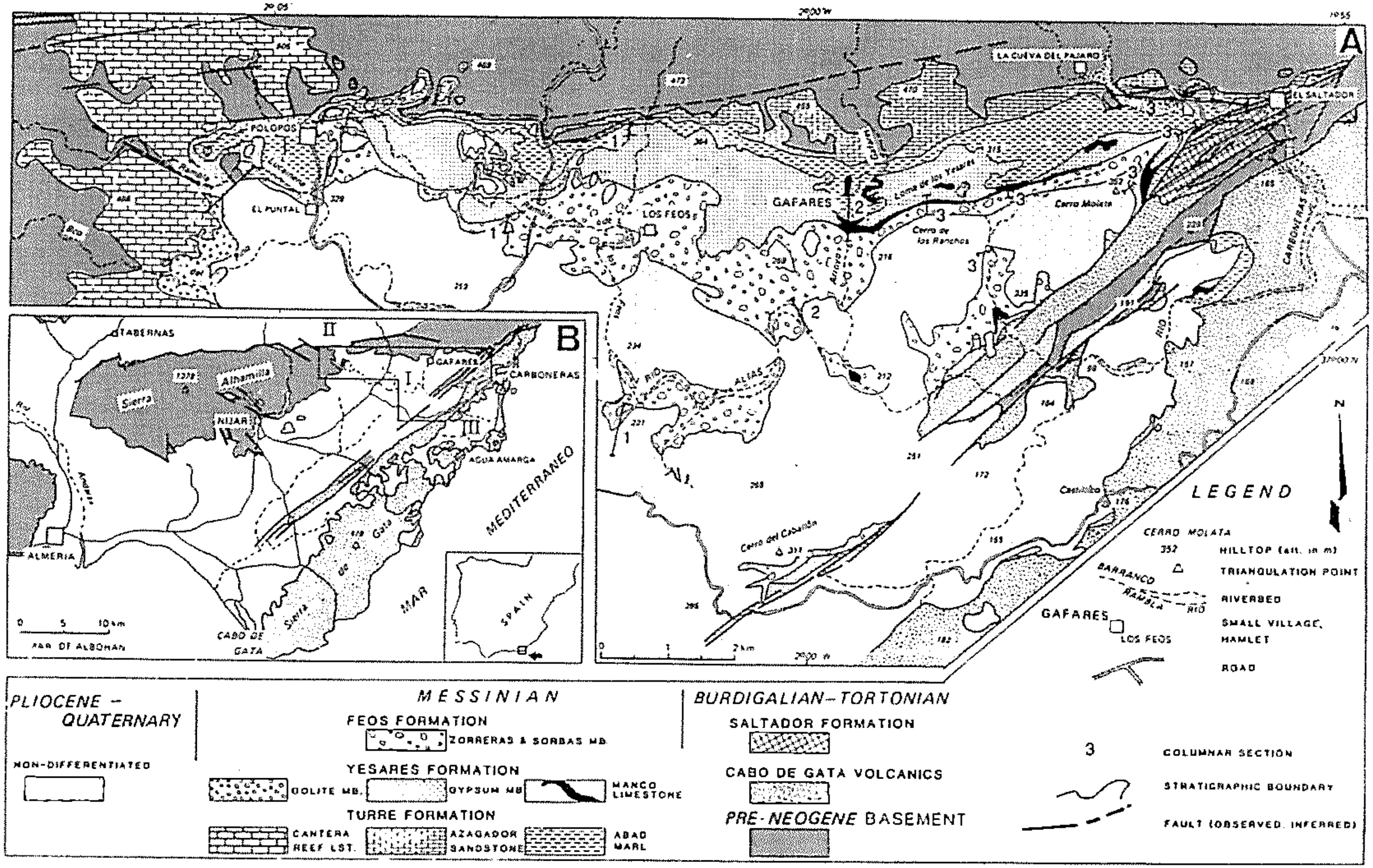


Fig. 2. (A) Geological map of the Northern Nijar Basin with location of measured composite sections 1-3, (B) Setting of the Northern Nijar Basin in the southern Almeria Province. I = Northern Nijar Basin, II = Sorbas Basin; III = Agua Amarga Basin.

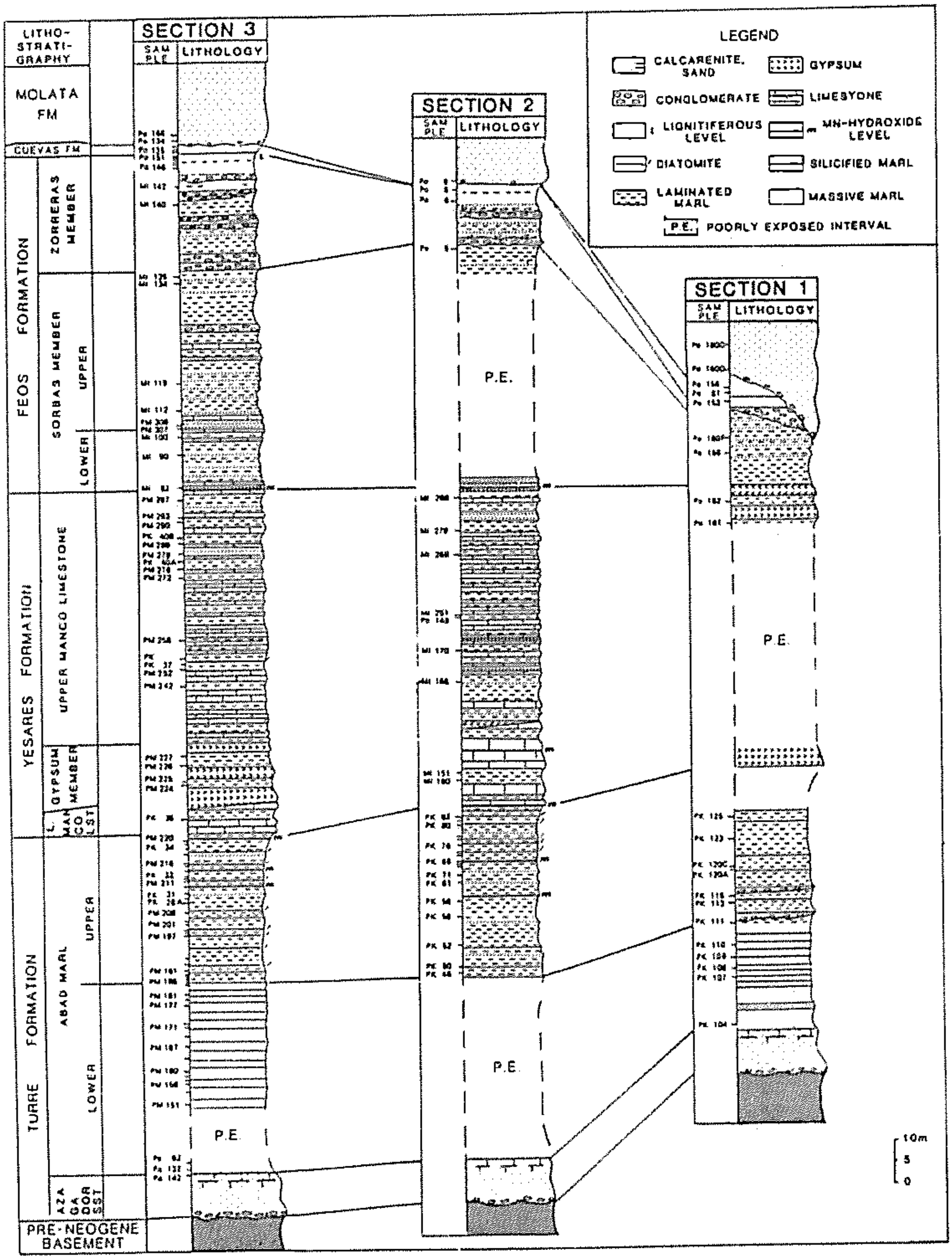


Fig. 3. Lithology, lithostratigraphic subdivision, correlation and sample position of measured composite Sections 1-3.

Assemblage A (samples Po 142-PM 183, Plate I).

Assemblage A is characterized by high frequencies of benthic foraminifers from both a diverse group (finely agglutinated species, fragile nodosariids, *Lenticulina spp, Planulina ariminensis,* 'small deep water taxa', *Cibicidoides pseudoungerianus*), and a first, distinct group of buliminaceans (*Uvigerina peregrina, Bulimina costata, B. aculeata* var. *minima, Bolivina arta, B. dilatata* and *B.subreticulata*) (App. IIa, c, Fig. 4; Plate I). A diverse planktonic foraminifer fauna (commonly about 70% of the foraminifera) and the presence of glauconite further characterize Zone A. This unit has three subzones.

Subassemblage A1 covers the lowermost sample (Po 142), in which *Cassidulina, Cibicidoides* and epiphytes dominate the benthic foraminiferal assemblage together with *Elphidium crispum.* In contrast to the relatively low foraminifer number, it has a high content in bryozoan and echinoid remains and sand-sized terrigenous debris.

An increase in frequency and diversity of the benthic foraminifera marks the base of **Subassemblage A2** (samples Po 132-PM 164), which is characterized by the presence of *Uvigerina striatissima,* finely agglutinated taxa and *Anomalinoides colligerus..*

Diversity decreases in the upper **Subassemblage A3** (samples PM 167-183), where *Cibicidoides pseudoungerianus* is dominant and *Bulimina costata* and *Uvigerina peregrina* reach high numbers, whereas characteristic elements of Assemblage B make their first appearance. A similar *C. pseudoungerianus* assemblage was found in the middle part of the Abad Marl of the adjacent Sorbas Basin (Troelstra et al. 1980; Chapter 4.3).

Plate 1

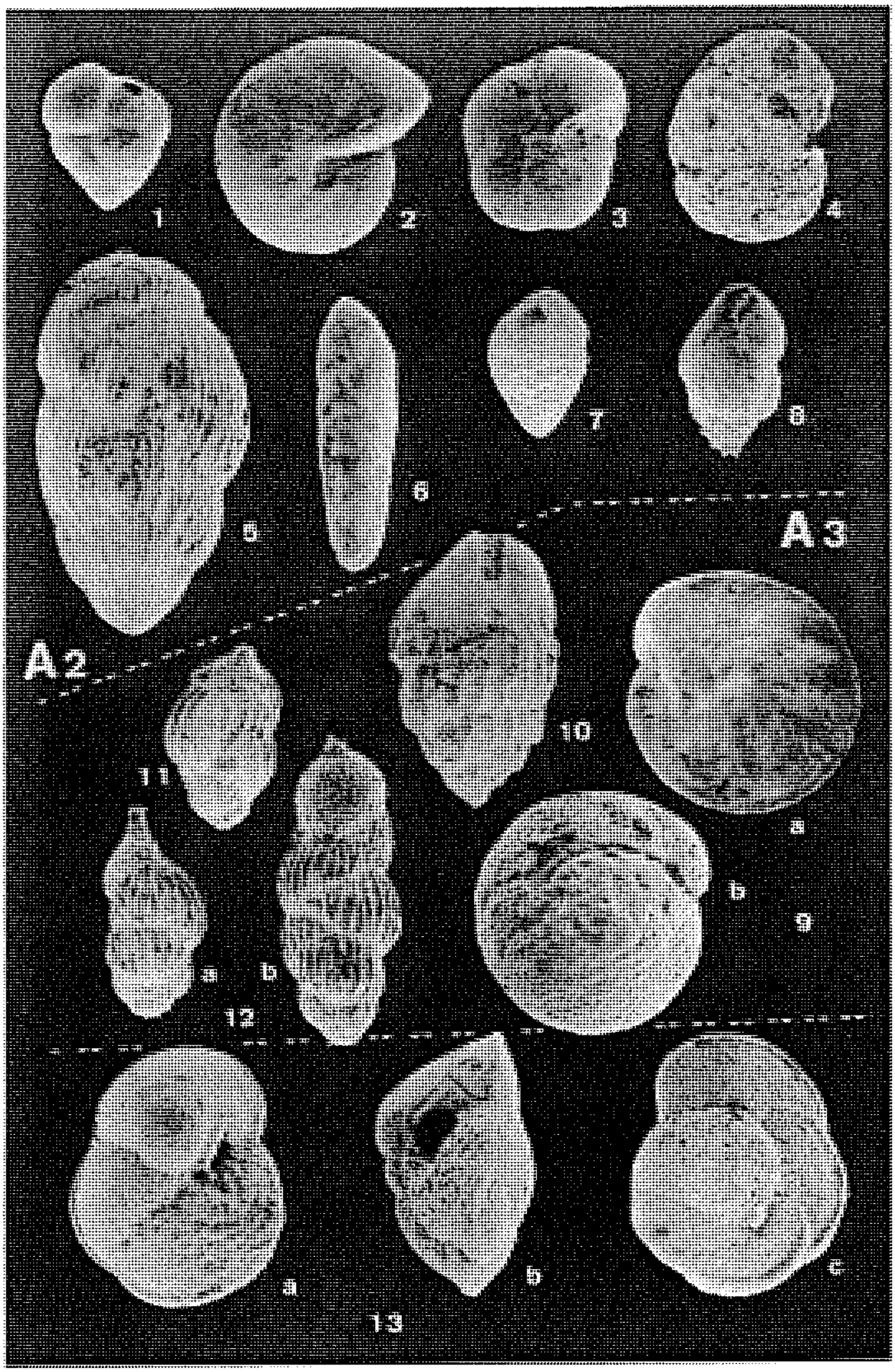
SEM micrographs of foraminifers characteristic for the Late Tortonian to early Messinian; all specimens x 70 except for 13 a-c (x 110).

Figs. 1-8, benthic species characteristic for the latest Tortonian to earliest Messinian "deep open marine" Assemblage A2.
Fig. 1. *Eggerella bradyi* (Cushman); sample PM 158, section 3.
Fig. 2. *Melonis soldanii* (D'Orbigny); sample PM 157, section 3.
Fig. 3. *Pullenia quinqueloba* (Reuss); sample PM 157, section 3.
Fig. 4. *Elphidium fichtellianum* (D'Orbigny); sample PK 49, section 2.
Fig. 5. *Uvigerina striatissima* Perconig; sample PM 164, section 3.
Fig. 6. *Bolivina arta* Macfayden; sample PM 157, section 3.
Fig. 9. *Bolivina dilatata* Reuss; sample PM 151, section 3.
Fig. 8. *Bulimina aculeata* D'Orbigny var. *minima* Tedeschi and Zanmatti; sample PM 158, section 3.

Fig. 9-11, benthic species characteristic for the early Messinian, marine, slightly oxygen-deficient Assemblage A3.
Fig. 9. *Cibicidoides pseudoungerianus* (Cushman); a. umbilical side, b. spiral side; sample PM 157, section 3.
Fig. 10. *Bulimina costata* D'Orbigny; sample PM 151, section 3.
Fig. 11. *Uvigerina peregrina* Cushman with costae on last chamber; sample PM 164, section 3.
Fig. 12. *Uvigerina peregrina* Cushman with pustulous (?and spinose) last chamber; a. short specimen, b. elongate specimen; sample PM 164, section 3.

Fig. 13, Plamktonic foraminifer characteristic for the early Messinian.
Fig. 13. *Globorotalia conomiozea* Kennett; a. umbilical side, b. side view, c. spiral side; sample PM 153, section 3.

Plate I



Assemblage B (samples PM 185-220; Plate II).

Assemblage B is first characterized by the consistent domination of a particular group of buliminaceans (*Bolivina spathulata*, accompanied by *B. dentellata*, *Bulimina elongata*, *Globobulimina pyrula*, *Rectuvigerina cylindrica* variants and *Hopkinsina bononiensis*). Additional shallow-water taxa (*Bolivina plicatella*, *Cancris auriculus*, *Valvulineria bradyana*, small 'epiphytes', *Ammonia* and *Elphidium* spp., *Florilus boueanum*, occasional miliolids) are often found (App. IIa-c, Fig. 4). The planktonic foraminifer assemblage similarly shows a low diversity with a tendency for different (groups of) taxa (*G. bulloides* s.l./*Globigerinella siphoniphera*, *Orbulina*/*Globigerinoides* spp, *Turborotalita quinqueloba*) to dominate in successive samples.

Small species are abundant in the low-diverse, but rich and well-preserved, foraminifer faunas. Besides, frequent fragile bryozoan remains, sponge spicules, fish scales

Plate II

SEM micrographs of foraminifers characteristic for the late early, middle and upper Messinian Assemblages B, C and D; all specimens x 70, except for 11a-c and 12a,b (x 110).

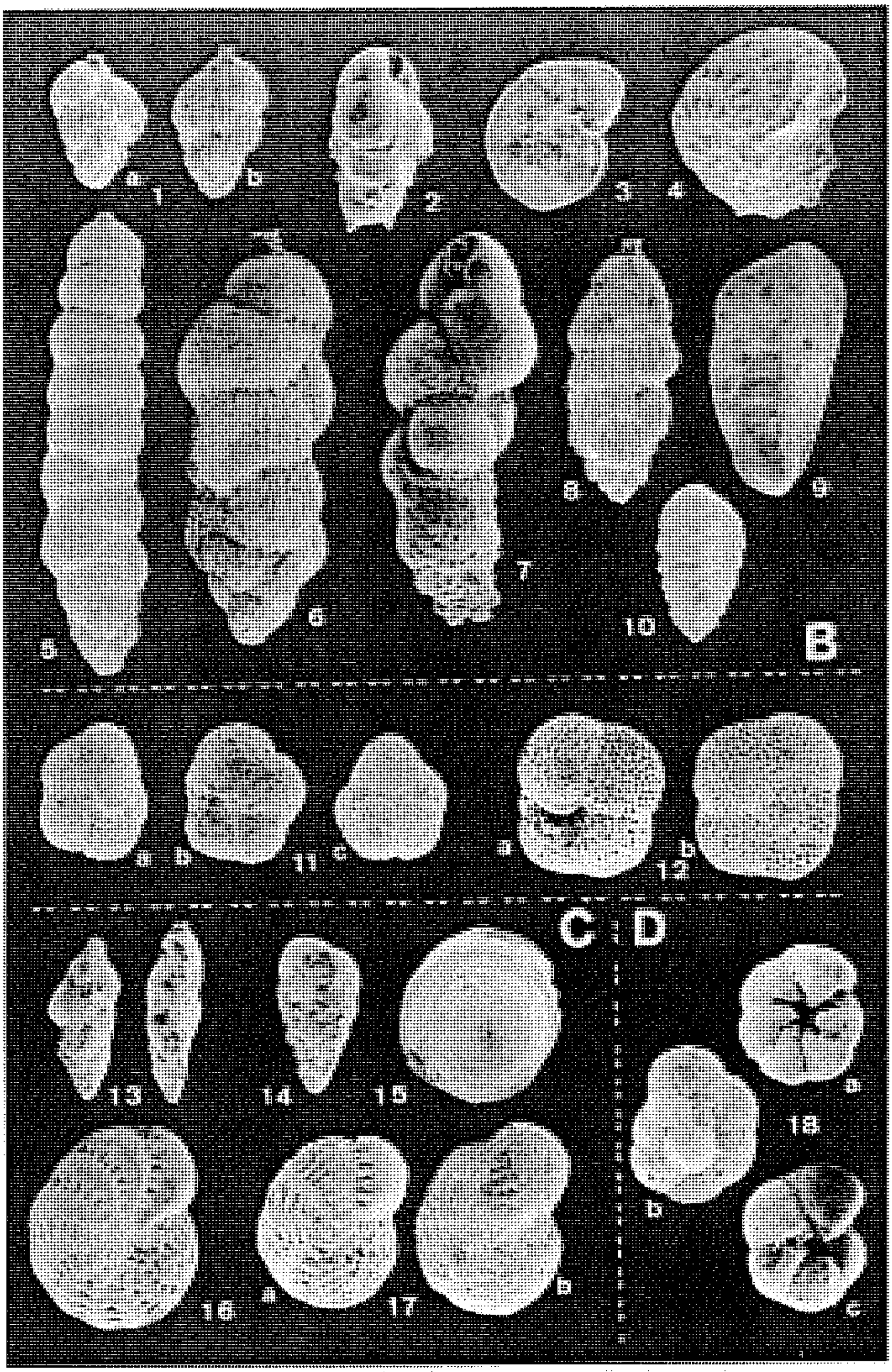
Figs. 1-10, benthic species characteristic for the late early to early middle Messinian, marine, low-oxygen Assemblage B:
Fig. 1. a,b. Small *Uvigerina* sp. cf. *U. peregrina* Cushman with costae on last chamber; sample PM 201, section 3.
Fig. 2. *Bulimina aculeata* D'Orbigny; sample PK 61, section 2.
Fig. 3. *Hanzawaia bouena* (D'Orbigny); sample PM 173, section 3.
Fig. 4. *Elphidium aculeatum* (D'Orbigny); sample PK 201, Section 3.
Fig. 5. *Rectuvigerina cylindrica* (D'Orbigny) var. *cylindrica* (Thomas); sample PK 56, section 2.
Fig. 6. *Rectuvigerina cylindrica* (D'Orbigny) var. *gaudryinoides* (Lipparini); sample PK 111, section 1.
Fig. 7. *Bulimina elongata* D'Orbigny; sample PM 201, section 3.
Fig. 8. *Hopkinsina bononiensis* (Fornasini); sample PK 111, section 1.
Fig. 9. *Bolivina spathulata* (Williamson); sample PK 56, section 2.
Fig. 10. *Bolivina dentellata* Tavani; sample PM 201, section 3.

Figs. 11, 12, characteristic planktonic foraminifera of the middle Messinian:
Fig. 11. *Turborotalita quinqueloba* (Natland); a: form with lip; b: most common form; c: atypical form, with coiling as in *Globigerina*; three specimens from sample PM 201, section 3.
Fig. 12. Dextral *Neogloboquadrina acostaensis* (Blow); a. umbilical side, b. spiral side; sample PK 58, section 2;

Figs. 13-15, 17. Benthic species characteristic for the late middle Messinian, restricted marine Assemblage C; the species of Fig. 16 is from the Recent.
Fig. 13. a, b. The small *Uvigerina* sp. A; sample MI 151, section 2.
Fig. 14. *Bolivina plicatella* Cushman; sample PK 52, section 2.
Fig. 15. *Asterigerina planorbis* D'Orbigny; sample PK 56, section 2.
Fig. 16. *Elphidium* sp. cf. *E. margaritaceum* Cushman; sample HP 88OO2, Salin de Gruissan (Aude, S. France).
Fig. 17. *Elphidium* sp. A; a. normally developed specimen; b. specimen with aberrant last chamber; sample MI 308, section 3.

Fig. 18. *Ammonia tepida* (Cushman), bentic foraminifer characteristic of the late Messinian 'continental' Assemblage D.; a. umbilical side, b. spiral side, c. umbilical side; three specimens from Sample MI 112, Section 3.

Plate II



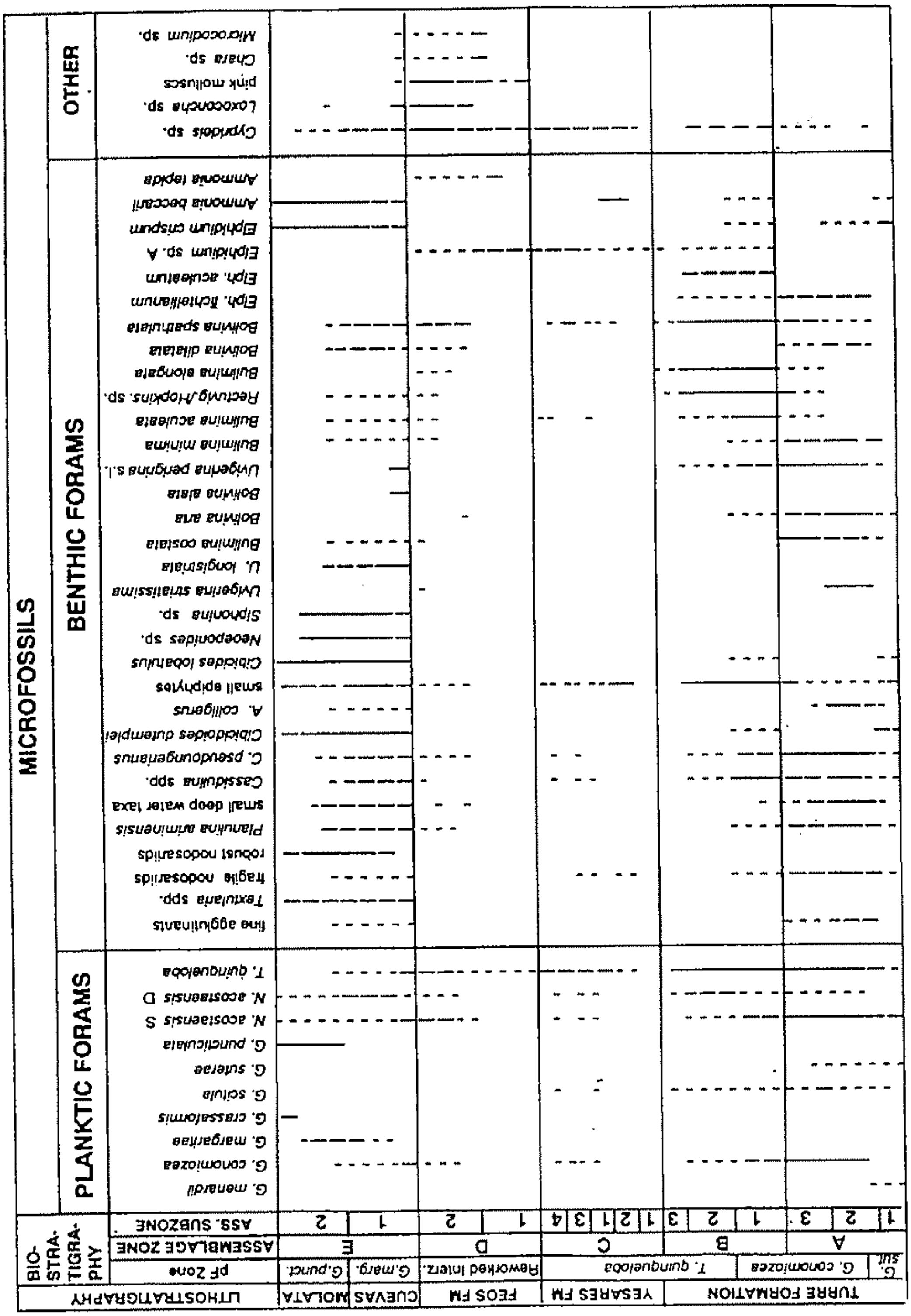


Fig. 4. Ranges of foraminiferal taxa with chronostratigraphic and environmental importance.

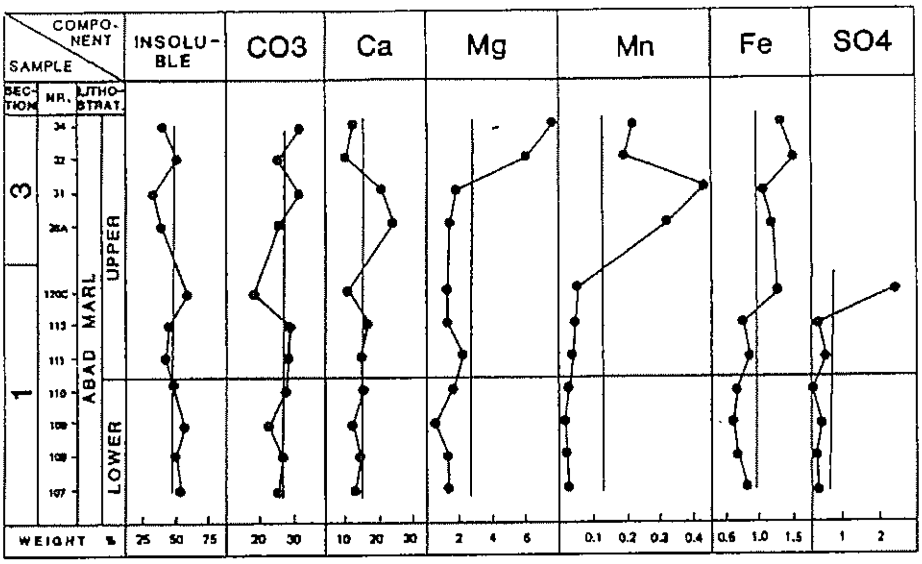


Fig. 5. Geochemical data on the Abad Marl. Note: higher Mg, Mn, Fe, and SO4 content for Upper Abad Marl, showing its higher dolomite content, especially in its top, and enrichment in metal-hydroxides and fine-crystalline gypsum.

and teeth, and, more scattered, diatom and thin pelecypod remains, are characteristic for this Assemblage. It coincides with the upper, laminated part of the Abad Marl which is rich in manganese and iron hydroxides and fine-crystalline gypsum, and has a relatively high carbonate content (App. I, IIa-c, Fig. 5; Van de Poel 1991).

The lower **SubassemblageB1** (samples PM 185-194) is characterized by a high frequency of *Cassidulina laevigata* and, more scattered, *Cibicidoides pseudoungerianus*, epiphytes, and globorotaliid and neogloboquadrinid planktonic foraminifers. *Globigerina bulloides* s.l. can reach high frequencies in this interval.

SubassemblageB 2 (samples PM 197-PK 31) has consistently low foraminifer diversity and strong fluctuations in P/B ratio.

An abrupt reduction in calcareous fauna determines the base of the upper **Subassemblage B3** (sample PM 210-220), which is further characterized by the occurrence of foraminifers with a glossy porcelaneous coating, masking their ornamentation.

AssemblageC (samples PK 35-MI 100, Plate II).

A final drop in diversity marks the lower boundary of Assemblage C, where the relatively larger benthic foraminifer species of Assemblage B virtually disappear and the remaining assemblage consists of a few small taxa (App. IIa-c; Fig. 4). *Elphidium* sp. A. is the most consistently present and, at certain levels, abundant. Besides, a few small bolivinids (especially *B. plicatella*) and epiphytes (*Asterigerina planorbis* and *Rosalina globularis*) are sometimes found. The planktonic foraminifer (*T. quinqueloba*) and ostracode fauna's (a.o. *Cyprideis* sp.) are also impoverished. Manganese hydroxide, gypsum and fish remains are the dominant, and often only, constituents in the washresidues.

Subassemblage C1 (samples PK 35-PM 246) is characterized an extreme paucity in microfossils, which partially may be ascribed to early dolomitization (Van de Poel 1991).

Subassemblage C2 in Section 2 (samples Ml 150-151), contains common *Ammonia beccarii*, small epiphytes and a few miliolids.

Subassemblage C3 (samples PM 252-PK 40A) represents a marine intercalation with common, well-preserved sponge spicules and fragile bryozoan remains. A sample from Section 2 (Po 143) contains *Uvigerina* sp. A and a relatively diverse small-sized planktonic foraminifer fauna.

Subassemblage C4 (sample PM 279-Ml 100) is again poor in microfossils but contains an interval with abundant fish remains (samples 289-40B, Section 3 and Ml 279, Section 2; De la Chapelle 1988).

Assemblage D (samples PM 307-Po 151, Plate II).

The appearance of pink lamellibranch and gastropod remains, *Chara* and *Microcodium* sp. marks the base of Assemblage D, in which well-preserved ostracodes of all mold sizes are the most common microfossils. *Ammonia tepida* is characteristic for the relatively rare autochthonous benthic foraminifer assemblage.

The lower **Subassemblage D1** (samples PM 307-Ml 135) still has a low microfossil diversity (*Cyprideis agrigentina, Elphidium* sp. A, *T. quinqueloba*) and contains manganese hydroxide and fish remains.

Subassemblage D 2 has typical brackish to fresh-water ostracodes (*Tyrrhenocythere pontica, Loxoconcha djaffarovi*) and is further characterized by a relatively poor, but diverse, marine foraminiferal assemblage and a high content in fine terrigenous debris and iron-hydroxides. Samples from its top contain lignite.

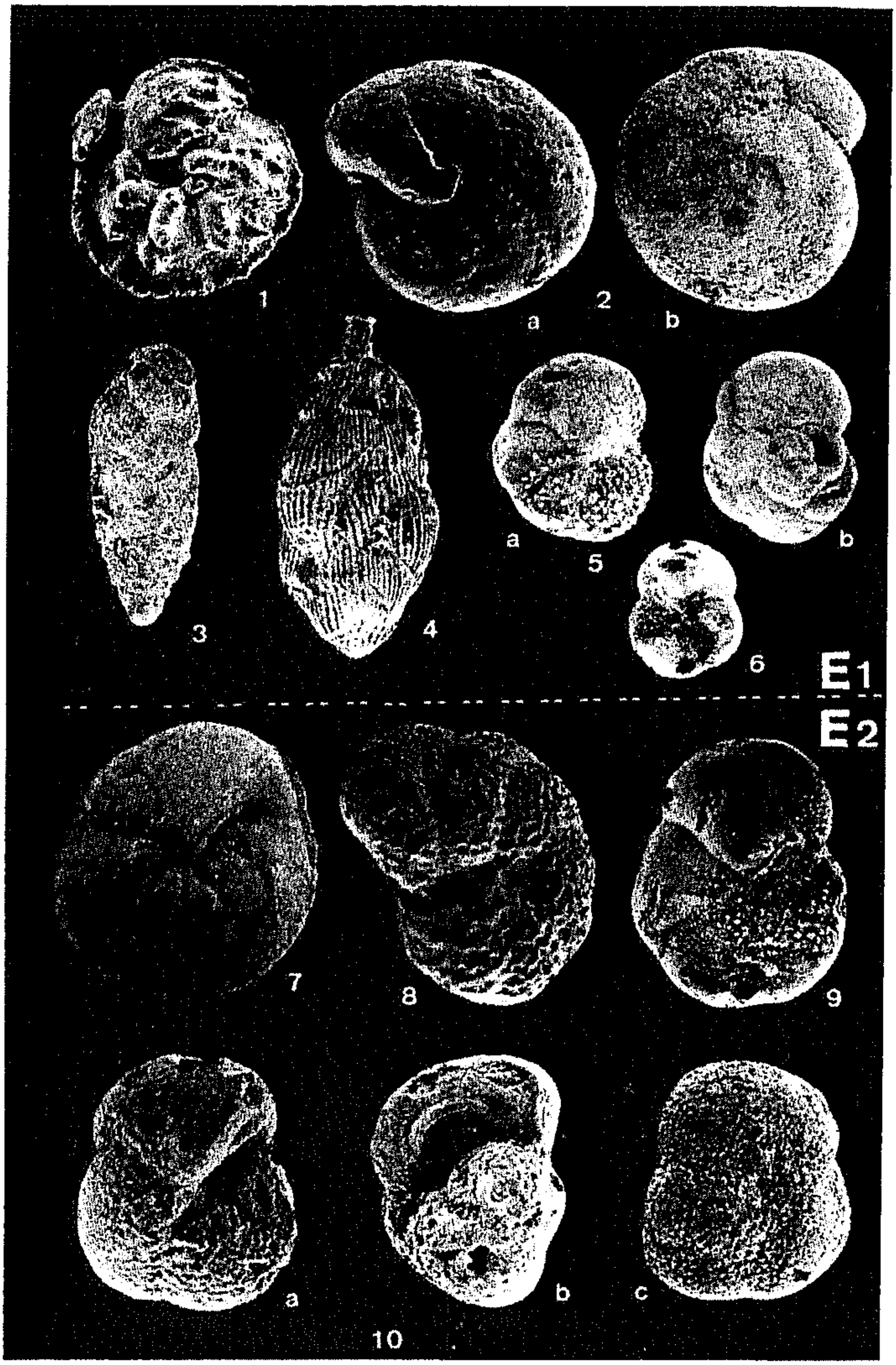
Plate III

SEM micrographs of characteristic, relatively shallow, open marine foraminifers from the Pliocene:

Figs 1-6, species characteristic of the early Pliocene, outer neritic Assemblage E1; 1-4 x 70; 5a, b-6 x 110.
Fig. 1. *Siphonina planoconvexa* Silvestri; sample 134, section 3.
Fig. 2. *Cibicidoides dutemplei* (D'Orbigny); a: umbilical side, b: spiral side; sample Po 135, section 3.
Fig. 4. *Bolivina alata* (Seguenza); sample PK 81, section 1.
Fig. 3. *Uvigerina longistriata* Perconig; sample 178, section 3.
Fig. 5a, b. The small *Globorotalia margaritae* Bolli and Bermudez forma *primitiva* Cita; a: umbilical side of sinistrally coiled specimen, b: dorsal side of dextrally coiled specimen.
Fig. 6. Umbilical side of particularly small, dextrally coiled specimen of *Globorotalia margaritae* Bolli and Bermudez forma *primitiva* Cita; sample Po 178, section 3.

Figs 7-10, species characteristic of the middle Pliocene, inner-neritic Assemblage E 2; 7 and 8 x 50; 9 and 10 x 100.
Fig. 7. *Neoeponides schreibersii* (D'Orbigny); sample Po 178, section 3.
Fig. 8. Coarse punctate type of *Cibicides lobatulus* (Walker and Jacob); sample Po 160D, section 1.
Fig. 9. *Globorotalia margaritae* Bolli and Bermudez forma *margaritae* Cita; sample Po 178, section 3.
Fig. 10. *Globorotalia puncticulata* Deshayes; a. umbilical side, b. side view, c. spiral side; two specimens from sample Po 160D, section 1.

Plate III



Reworking of the marine foraminifer fauna is first suggested by its mixed nature (diverse planktonic foraminifers, characteristic buliminaceans from Assemblage B and benthic species characteristic of Assemblage A), the occurrence of brown-coloured and damaged individuals, the presence of *Globorotalia conomiozea* and sinistral *N. acostaensis* which are characteristic for the earliest Messinian (both in Mediterranean and in nearby marine Atlantic sections: Bossio et al. 1977, Geerlings & Van de Poel 1979, Hodell et al. 1989), and the occurrence of early Messinian reef limestone fragments in associated river conglomerates (Van de Poel 1991). The non-marine ostracodes are well preserved.

AssemblageE (samples Po 152-178; Plate III).

The base of this Assemblage is marked by the reappearance of a rich and diverse benthic foraminifer assemblage, characterized by robust *Textularia* spp. and nodosariids, *Siphonina planoconvexa, Neoeponides schreibersii, Uvigerina longistriata,* and relatively common robust rotaliids. The abundance of characteristic lower Messinian (Assemblage A and B) taxa in this Assemblage is ascribed to reworking, indicated by the presence of typical early Messinian planktonic foraminifers (*Globorotalia conomiozea,* sinistrally coiled *N. acostaensis*) and continental Assemblage D markers (ostracodes, *Chara* and *Microcodium* remains). Preservation, abundance and diversity of the autochthonous planktonic foraminifers is moderate. The return of glauconite and common echinoid remains further marks the base of Assemblage E.

The lower **SubassemblageE1** is characterized by the abundance of *Cibicidoides dutemplei* and the relatively common occurrence of buliminaceans. A large part of the latter is considered reworked, but *Bolivina alata* only occurs in samples from this Zone (PK 81 and Po 155, Section 1).

Diversity again decreases in the upper Subassemblage E2 (samples Po 134-166), where *Cibicides lobatulus* is the most abundant benthic species, and *Globigerinoides* dominates the planktonic assemblage.

PLANKTONIC FORAMINIFER ZONES.

The following local planktonic foraminifer subdivision can be compared with the Mediterranean "standard" Zonation of Kastens, Mascle et al. (1990) (Fig. 6):

***Globorotalia suterae* Zone.**

Globorotalia suterae and *G. menardii* occur in the lowermost samples of Sections 1 and 3, where *Globorotalia conomiozea* is absent. Such an assemblage is typical for the uppermost Tortonian *Globorotalia suterae* Subzone of the *Globigerinoides obliquus extremus* Zone of Glaçon et al. (1990).

***Globorotalia conomiozea* Zone.**

From the first appearance of the nominate species to the change in coiling direction of *N. acostaensis* from predominantly sinistral to dextral (Geerlings & Van de Poel 1979, Iaccarino 1985, Glaçon et al. 1990).

Turborotalita quinqueloba Zone.

From the coiling change of *N. acostaensis* up to the top of the interval in which, consistently, the nominate species is the dominant form. The term *T. multiloba* Zone for this interval (Colalongo et al. 1979), is not used, since the endemic *T. multiloba* Romeo is very rare in the investigated samples. The same has been observed in the adjacent Sorbas Basin (Geerlings 1977a, b).

A **Barren Interzone** is recognized for the interval in which planktonic foraminifers are either absent or extremely scarce.

Reworked Interzone.

This interval shows an upwards increase in frequency and diversity of the planktonic foraminifer fauna, which may be well preserved, but is mixed, as indicated by the presence of *G. conomiozea* and sinistrally coiled *N. acostaensis,* and is considered reworked. Reworking is further indicated by the presence of damaged shells, brown individuals and abundant fine terrigenous debris (see also under Assemblage D).

Globorotalia margaritae **Zone.**

From an abrupt increase in frequency and diversity of planktonic foraminifers, accompanied by other marine microfossils, up to the first appearance of *Globorotalia puncticulata* (App. IIa-c; Fig.4). A few, poorly preserved individuals of the nominate taxon (predominantly the small *G. margaritae primitiva* Cita) are present in most of the samples. *G; margaritae* is absent from the lowermost samples of this interval in Section 3. Although a few *Sphaeroidinellopsis* sp. have been observed the material does not allow for a confirmation of the presence of the earliest Pliocene *Sphaeroidinellopsis* Acme Zone (MPL1) of the central Mediterranean (Cita 1975b, McKenzie & Sprovieri 1990). Its presence can be demonstrated at the base of the Cuevas Formation of the adjacent Vera Basin ('Zone du Passage' of Montenat et al. 1976).

Globorotalia puncticulata **Zone.**

From the first appearance of the nominate species up to the top of the section. This Zone contains advanced *G. margaritae* forms. The youngest sample (Po 160 C, Section 1), which still contains *G. puncticulata,* but *Globorotalia crassaformis* instead of *G. margaritae* may represent Zone MPl 4 of Cita (1975b) (Figure 6).

ENVIRONMENTAL INTERPRETATION AND AGE OF THE EVENTS

The recognized succession of asssemblages serves to interpret the evolution of some paleoenvironmental factors (waterdepth, salinity and oxygen content), whereas the planktonic foraminifer zonation allows for an age assignment to the events (Figs 6 & 7).

AGE.

The main planktonic foraminifer events, and therewith the changes in facies, can be dated in terms of the timescale of Berggren et al. (1985, slightly modified in Figs 6 and 7, left part).

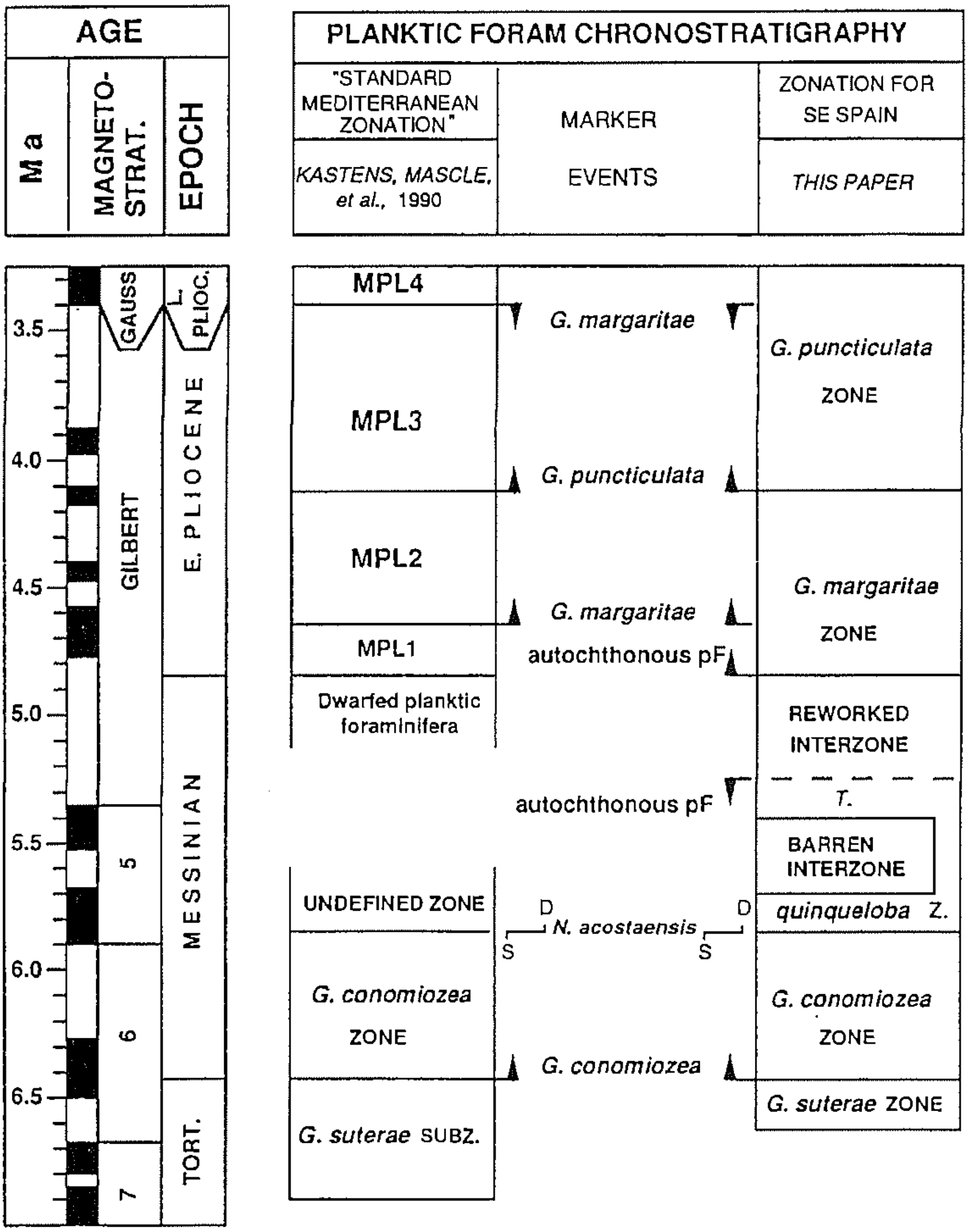


Fig. 6. Comparison of the local planktic foraminiferal zonation and events with the 'Standard Mediterranean Zonation', and their correlation with the timescale of Berggren et al. (1985, slightly modified after Zijderveld et al., 1986 and Channell et al., 1990).

The first appearance of *G. conomiozea* in the Mediterranean occurs within the Chron 6 Normal Zone and has an age of 6.43 Ma (Channell et al. 1990). The boundary between Subassemblages A1 and A2, which coincides with the lithologic boundary between Azagador Sandstone and Abad Marl, is slightly older, being situated in the upper part of the *G. suterae* Zone (total age ~6.90-6.43 Ma; Glaçon et al. 1990), and is estimated at 6.5 Ma.

The sinistral to dextral coiling change of *N. acostaensis* is estimated at 5.85 Ma since it approximates, according to Bossio et al. (1977), the first appearance of *Amaurolithus tricorniculatus* which falls in the lower normal interval of Chron 5 (Haq et al. 1982), and slightly predates the first appearance of *Pulleniatina primalis* , which also takes place in the lower normal of Chron 5 at 5.8 Ma (Saito et al. 1975, Van Gorsel & Troelstra 1981, Berggren et al. 1985).

The next three boundary ages are calculated using the above ages and assuming constant sediment accumulation rates for the Abad Marl.

The age of initiation of faunal impoverishment, marked by the base of the A3 Subassemblage, which characterizes the top of the Lower Abad Marl, has been calculated at 6.10 Ma. The same age is given by Hodell et al. (1989) to a shift in $\partial^{18}O$ in the Bou Regreg section (NW Maroc), which they correlate with the onset of diatomaceous 'tripoli' sedimentation in the Sicilian Messinian type area.

The major change in fossil and lithologic composition which marks the base of Assemblage B and of the Upper Abad Marl, respectively, has been calculated to be at 6.0 Ma. This is close to the age calculated for the base of Sicilian Tripoli Formation by Schrader & Gersonde (1978) and Gersonde & Schrader (1984).

The record of Section 3 suggests a gradual onset of 'tripoli' sedimentation, the Upper Abad Marl correlating with the Sicilian Tripoli Formation on basis of lithology, stratigraphic position and the coiling change of *N. acostaensis* in the middle of both formations (Van de Poel 1991). The "acme" of *G. bulloides,* which preceedes the *N. acostaensis* coiling change, reflects eutrophic conditions (Reiss & Hottinger 1984, Troelstra & Kroon 1989, Barmawidjaja et al. 1989). In Section 3 it is situated at the level where the first common diatomites have been observed.

The age of the boundary between Assemblages B and C, which coincides with the boundary between Abad Marl and Yesares evaporites, is calculated at 5.6 Ma. This is close to the 5.7 Ma calculated by Gersonde & Schrader (1984) on basis of diatom correlations for the boundary between Tripoli Formation and the Messinian 'Lower Evaporites' of the Mediterranean, which is commonly used in Messinian 'scenarios' (McKenzie et al. 1985, Müller & Hsü 1987). It confirms a former stratigraphic contention of the approximate age-equivalence of the initiation of Yesares and central Mediterranean evaporite deposition (Van de Poel 1991).

The abrupt boundary between the Feos and Cuevas Formations (boundary between Assemblages D and E) is equated to the base of the Pliocene in the Mediterranean, which falls in the top of the lower reversed interval of the Gilbert Chron, just below the Thvera Subchron and has an age of ca. 4.85 Ma (Zijderveld et al. 1986, Channell et al. 1990). As recorded in the *Sphaeroidinellopsis* Acme, it is contemporaneous in the Vera Basin of SE Spain and in the central Mediterranean, but needs further detailed investigation in the Carboneras-Nijar area.

Ages for facies events within the interval between the two latter datum-levels (i.e. within the Yesares and Feos Formations/Assemblages C-D) are problematical due to the lack of biostratigraphic markers.

The marine intercalation in the higher part of the Yesares Formation (Subassemblage C3) compares with a similar level near the top of the Yesares Formation of the adjacent Sorbas Basin (Montenat et al. 1980, Ott d'Estevou 1980) and is possibly correlatable to a marine interval in the Fortuna Basin (SE Spain) for which an age of ca. 5.35 Ma has been suggested by Müller & Schrader (1989).

The youngest Messinian deposits then entirely fall in the lower reversed period of the Gilbert Chron as is the case in the central Mediterranean (Channell et al. 1990). The brackish Assemblage D of the Feos Formation is characteristic of the Mediterranean 'lago mare' (Ruggieri & Spovieri 1976, Colalongo et al. 1978, Cita et al. 1978) and correlates with a strong glacial maximum found in the world oceans in the lower reversed of the Gilbert Chron (Van Gorsel & Troelstra 1981, Müller & Hsü 1987; McKenzie et al. 1988).

G. puncticulata first appears in the Mediterranean at the end of the Gilbert normal B chron at 4.13 Ma (Zijderveld et al. 1987, Rio et al. 1990). The *G. puncticulata* base has been found by Völk (1967) just below the top of the Cuevas Formation in the Vera Basin, but in the Carboneras area much of the Cuevas Formation has probably been eroded. Since the Molata Formation (and Subassemblage E2) contains *G. puncticulata,* its transgressive base is somewhat younger than this event and may coincide with the sequence boundary at 3.8 Ma of Haq et al. (1987).

G. margaritae has its last occurrence close to the top of the Gilbert Chron at 3.5 Ma (Berggren et al. 1985, Rio et al. 1990). Our youngest sample, from the upper part of the Molata Formation, which contains no longer *G. margaritae,* but still *G. puncticulata,* should have an earliest Late Pliocene age between 3.5 and 3.3 Ma (figure 10 of Rio et al. 1990).

PALEOENVIRONMENTS (Fig. 7).

At the end of the Tortonian, the near-shore environment recorded in the main part of the Azagador Sandstone (Van de Poel 1991), was rapidly replaced by the open marine conditions indicated by the diversity of benthic and planktonic foraminifers and the presence of glauconite in **Assemblage A**.

The abundance of *Amphicoryna scalaris, Lenticulina* spp., *Cibicidoides pseudoungerianus, Planulina ariminensis, Pullenia, Gyroidina* and *Melonis* spp., the presence of *Bigenerina nodosaria* and *Eggerella bradyi,* and the virtual absence of *Cibicidoides dutemplei* and other typical shelf-taxa, suggest an upper bathyal waterdepth of 300-400 m for **Subassemblage A2** (cf. Pujos 1976, Jorissen 1988, Van Marle,1988).

Subassemblage A3, with its first reduction in benthic diversity, records the initiation of aberrant bottom conditions towards the end of the earliest Messinian. Its increase in importance of buliminaceans indicates increased nutrient and/or lowered oxygen levels at the sea bottom (Van der Zwaan 1982, Reiss & Hottinger 1984, Troelstra 1989). Yet, non-buliminacean benthic foraminifera remain present and the abundance of

Cibicidoides pseudoungerianus, Bulimina costata and *Uvigerina peregrina* suggests that oxygen depletion did not reach extreme values (Van der Zwaan 1982, Jonkers 1984, Jorissen 1987). The fauna compares with those of the margins of oxygen minimum zones in upwelling areas (Harman 1964, Ingle et al. 1980, Lutze 1986, Van Marle 1988, Hermelin & Shimmield 1990). The observed fluctuations in relative frequencies of the buliminaceans reflect fluctuations in oxygen content of the bottom waters.

The presence of 'small deep water species' and the frequency of *Bulimina costata, Planulina ariminensis, Bolivina spathulata* and *Cibicidoides pseudoungerianus* (rather than *C. bradyi* (Trauth)), suggest a 250-350 m waterdepth for this interval (Pujos 1976, Ingle et al. 1980, Jorissen 1988, Kouyoumontzakis 1987, Van Marle 1988).

The dominance of buliminaceans in **Assemblage B** reflects the high nutrient budget in the Mediterranean during the latest early Messinian (Van der Zwaan 1982). The abundance of small-sized species also is indicative of a high nutrient supply (Lankford 1959, Phleger & Soutar,1973). Strong oxygen-deficiency in the lower part of the water-column is further attested to by the common preservation of lamination (below 0.3 mg/l: Harman 1964), the high quality of preservation of the foraminiferal faunas (Phleger & Soutar 1975), and of fish and plant remains (Van de Poel 1991), whereas the occurrence of manganese-hydroxide levels also suggests 'suboxic' conditions at the sediment-water interface (Van de Poel & Klaver 1989). *B. spathulata* and comparable forms seem to avoid extreme oxygen deficiency (less than 0.1 mg/l) (Harman 1964, Jorissen 1988). Consequently, extremes were not reached during Assemblage B deposition.

The presence of *Rectuvigerina cylindrica* confirms the upwelling of deep Atlantic water into the Mediterranean during the latest early Messinian (Lutze 1986). Abundance of *B. spathulata* or comparable forms (i.e. *B. argentea* Cushman, *B.* aff. *rankini* Kleinpell, *B. ordinaria* Phleger and Parker) and/or the *Bulimina aculeata* plexus (*B. aculeata, Bulimina* sp. 1 Hermelin and Shimmield, *B. marginata-denudata, B. elongata*) is reported from both, oxygen minimum zones in upwelling areas (Harman 1964, Lutze 1979, Ingle et al. 1980, Troelstra et al. 1990, Hermelin & Shimmield 1990) and semi-enclosed marine basins (Reiss & Hottinger 1984, Jorissen 1987, Barmawidjaja et al. 1989). Both taxa have some tolerance to increased salinities (Reiss & Hottinger 1984, Debenay et al. 1987).

Aberrant conditions also prevailed in the higher part of the watercolumn as indicated by a decrease in diversity of the planktonic foraminifer faunas. Abundance of *G. bulloides* and *N. acostaensis* indicates high nutrient levels, whereas the abundance of *G. conomiozea* (the fossil equivalent of *G. menardii*) is suggestive of upwelling conditions (Van der Zwaan 1982, Troelstra & Kroon 1989). These species are most frequent in the older part of Assemblage B, while *T. quinqueloba*, globigerinoids (in particular *G. obliquus*) and *Orbulina* spp. dominate its younger part. Of these, only *T. quinqueloba* is sometimes seen to thrive particularly well in high-productive areas (pers. comm. G.J. Brummer and S.R. Troelstra). *T. quinqueloba* and the modern equivalent of *G. obliquus, G. ruber*, are surface-dwellers and together with *Orbulina* they are all three reputated for their tolerance to a relatively wide range of salinities (Ganssen & Troelstra 1987, Kroon et al. 1989, Bijma et al. 1990). Müller & Hsü (1987) have recorded a decrease in planktonic $\partial^{18}O$ in the top of th e Abad Marl of the adjacent Sorbas Basin, which was interpreted as pôssibly reflecting a greater fresh-water influence. However, the persistance of planktonic foraminifers in itself excludes extreme salinity variations in the surface waters.
The composition and, at levels, abundance of the 'shallow water component' of the benthic foraminifer fauna, indicates a well-vegetated environment with either normal marine salinity or a tendency to hyposalinity in the uppermost water levels (Lankford 1959, Murray 1973, Sen Gupta & Schafer 1973, Pujos 1976, references quoted in Van der Zwaan 1982, Zaninetti 1982, Coppa 1987, Debenay et al. 1987, Jorissen 1987, Moodley 1990, Culver 1990).

Alltogether, the foraminifer facies distribution of Assemblage B is in accordance with oxygen isotope data for the adjacent Sorbas Basin and basins in the central and eastern Mediterranean (Müller & Hsü 1987, Van der Zwaan 1982), which indicate the existence of a relatively strong salinity gradient during this interval.

Subassemblage B1 still has levels with somewhat more diverse benthic and planktic foraminifer faunas and shows fluctuations in the Fe and Mg content (Fig. 5). Levels with common *C. pseudoungerianus* and *Uvigerina* cf. *peregrina* and the general frequency of *Cassidulina* spp. suggest relatively better oxygenated conditions (Van der Zwaan 1982, Jonkers 1984, Barmawidjaja et al. 1989). This interval records a fluctuating shift towards the consistent conditions with low-oxygen and somewhat deviating salinities reflected by **Assemblage B2.**

The paucity in calcareous fossils in **Subassemblage B3** is probably largely the result of secondary dolomitization (see also Mg-curve of Fig. 5), which can give rise to dissolution of calcareous tests (Kelts and McKenzie 1981). In the adjacent Sorbas Basin, where dolomitization of the uppermost Abad Marl is not evident, common, normally preserved calcareous microfossils remain present in this interval (Chapter 4.3).

The disappearance of all 'deep water' indicators near the base of Assemblage B, and the concomitant appearance of a number of shallow-water species, suggest a shoaling trend in the Abad Marl (Troelstra et al. 1980, Chapter 4.3). The sudden abundance of very shallow water indicators (small specimens of *Elphidium* spp., *Ammonia*, miliolids, 'epiphytes', shallow water ostracodes) in the laminites at the base of Assemblage B is partially due to displacement by turbidity currents: large, robust specimens of the same taxa occur in samples from turbidite beds. Floating plant material, such as the large seagrass fragment found in the Upper Abad Marl of Section 1 (Chapter 3.2: Plate IIc), probably further formed an important transport vehicle for epiphytes (Van der Zwaan 1982, Jorissen 1987, Troelstra 1989, p. 159). *Bolivina plicatella* (which may also be epiphytic: Van der Zwaan 1982), *Hopkinsina bononiensis* and *Bulimina elongata* also could have been transported since they are commonly reported from modern, shallow, often slightly hyposaline waters (Murray 1973, Pujos 1976, Haake 1977, references in Van der Zwaan 1982, Zaninetti 1982, Lutze 1986, Coppa 1987, Debenay et al. 1987, Moodley 1990, Culver 1990). Waterdepth estimates have to be made on the remaining benthic elements. The presence of *U. peregrina*, *C. pseudoungerianus*, and some *P. ariminensis* and 'small deep water species' in the base of this Assemblage suggests waterdepths of at least 150m for this interval (Pujos 1976, Jorissen 1988, Van Marle 1988). *Hanzawaia boueana*, *Florilus boueanum* and *Valvulineria bradyana* may well be autochthonous in Assemblage B, since they are relatively common and have been reported from modern environments dominated by buliminaceans. They are reported as common in waterdepths up to 200 m and sometimes more (Pujos 1976, Ingle et al. 1980, Van der Zwaan 1982, Jorissen 1988). Abundance of *Bolivina spathulata*, the most common species of Assemblage B, is characteristic of the shelf edge and upper slope (Murray 1973, Pujos 1976, Jorissen 1988). Comparison of the uvigerinid species of Assemblage B with those of the modern central-eastern Atlantic, where *Hopkinsina bononiensis* has a lower depth limit of 150 m and *Rectuvigerina cylindrica* an upper depth limit of 150 m and its maximum between 250 and 500 m (Lutze 1986), indicates a waterdepth below 150 m for the entire Assemblage B. Alltogether, 150 to 250 m waterdepth seems a reasonable conclusion. This depth fits the geometry of reconstructed topographic differences between Upper Abad Marls and contemporaneous marginal reef deposits (such as those at Polopos: Fig. 2; Troelstra et al. 1980, Van de Poel et al. 1984, Van de Poel 1991).

Assemblage C, recognized in interbedded marls of the Yesares and basal Feos Formation, shows general faunal paucity, whereby completely barren intervals are probably the combined result of an extreme original environment and secondary dolomitization (Van de Poel 1991). The strongly reduced diversity in Assemblage C, and the fact that *Elphidium*

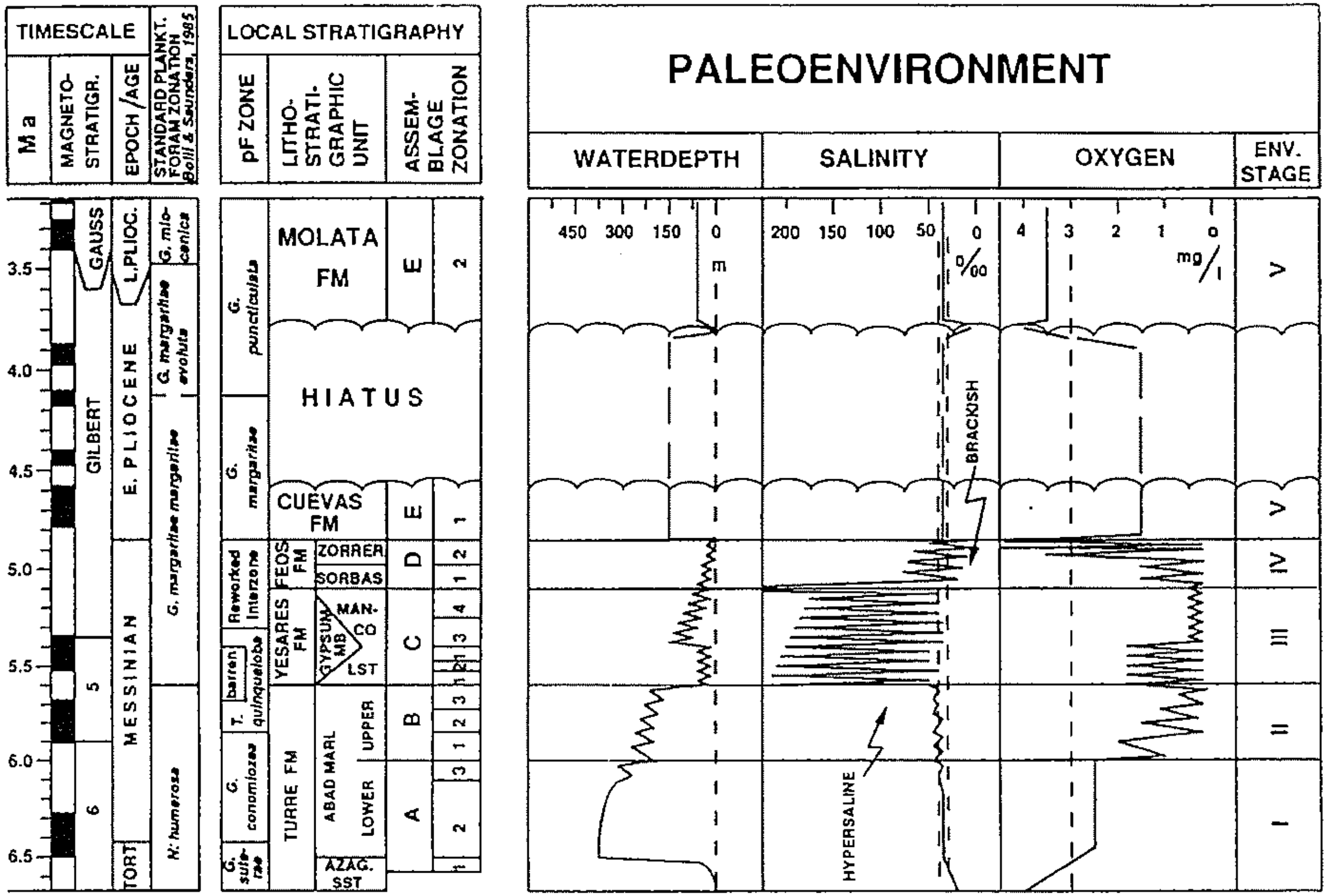


Fig. 7. Estimated depth, salinity and oxygen-content of the bottom waters for the centre of the late Tortonian-middle Pliocene Northern Nijar Basin.

and *Cyprideis* (two genera known to thrive under hyposaline and hypersaline conditions: Truc 1980, De Deckker et al. 1988) are relatively common, further indicate that the open marine connection had become considerably obstructed. The association of Assemblage C with gypsum suggests continuously hypersaline conditions, but it is peculiar that miliolids and *Ammonia beccarii* are relatively rare as they are known to thrive in shallow hypersaline environments, commonly in company of *Elphidium* and *Cyprideis* (Murray 1973, Zaninetti 1982, Debenay et al. 1987, personal observations on samples from the natural salt works of Gruissan, southern France). This can be explained by assuming a relatively deep water environment for the main part of Assemblage C.

Subassemblage C2 contains a benthic foraminifer fauna (common small *Ammonia*, accompanied by diverse small epiphytes, some miliolids, *Elphidium*, *Bolivina plicatella* and *Fursenkoina*) which compares with faunas described from modern shallow lagoonal environments (Debenay et al. 1987, Culver,1990).

The relatively constant open marine signal (*T. quinqueloba*, scattered marine benthic foraminifera, sponge spicules and bryozoan remains) in Assemblage C, can often be linked to the relative abundance of fine terrigenous clastics, and hence to reworking. The richer, internally consistent fossil assemblages of **Subassemblage C3**, which sometimes include abundant, well-preserved remains of marine sponges, are considered to be the result of temporary marine incursions. The reconstructed topographic heigth difference between the deposits in which Assemblage C occurs and contemporaneous oolite-rich beds from the basin margin (exposed near Polopos, Fig. 3.11), indicates, at least temporary, considerable waterdepths, comparable to those of Assemblage B, in the centre of the basin (Van de Poel 1991, fig. 9).

Oxygen-content of the bottom waters was low as attested by the abundance of lamination. *Elphidium*, *Cyprideis* and *Ammonia* are taxa that can survive low oxygen levels that are common in lagoons or estuaria, where they proliferate (Lutze 1965, Moodley 1990, personal observations in the 'Salin de Gruissan').

Assemblage C attests to the transition from the open marine environment of the early Messinian to the late Messinian 'continental' conditions recorded in **Assemblage D**.

The most common microfossils of **Subassemblage D1** (*Cyprideis* and *Ammonia tepida*) can again be interpreted as reflecting either a brackish or hypersaline environment. A relatively high salinity level may have been maintained by the reworking of evaporites from the middle Messinian Yesares Formation (Van de Poel 1991). The presence of typical brackish 'lago mare' mollusks and of *Chara* and *Microcodium*, on the other hand, suggest an increased influx of fresh water in the basin in comparison with the underlying interval. Assemblage D1 may well have originated under rapidly fluctuating, 'schizohaline' conditions (Colalongo et al. 1978). Sedimentological features and reconstruction of the basin topography suggest that several tens of metres waterdepth were still commonly attained during this time-interval (Roep & Van Harten 1979, Van de Poel 1991).

The presence of low-salinity ostracode species and lignite fragments in **Assemblage D2** record a further freshening and shallowing of the basin towards the end of the Messinian, which is corroborated by the sedimentary facies (Roep & Van Harten 1979, Van de Poel 1991). The attenuation of lamination, and the presence of limonite, burrowing and abundant mollusk remains indicate better oxygenation of the basin at the same time.

A sudden return to open marine conditions is marked by the abrupt increase in microfossils and by the reappearance of echinoids and glauconite, at the base of Assemblage E.

The relative frequency of planktonic foraminifers and the dominance of *Cibicidoides dutemplei* indicate outer shelf waterdepths for Subassemblage E1 (Kafescioglu 1975, Van Marle 1988). Depth may have been somewhat greater at the location of Section 1, where *Cibicidoides pseudoungerianus* and *Bolivina alata* are common.

The local angular unconformity at the base of the Molata Formation and the poor development of the lowest Pliocene in the Carboneras area, attest to a period of uplift and erosion (Van de Poel et al. 1984, Van de Poel 1991; Figs 6, 7), after which the shallow marine conditions of middle Pliocene Subassemblage E2, in which *Cibicides lobatulus* and *Elphidium crispum* dominate, were installed.

CONCLUSION AND DISCUSSION

Five distinct biostratigraphic units of the Carboneras-Nijar Basin represent the latest Tortonian to early Pliocene succession of relatively deep, open marine (I), relatively shallow marine, eutrophic (II), relatively shallow to shallow, very restricted marine (III), brackish, continental (IV), and renewed open marine, but shallow (V) 'Environmental stages' (Fig. 7).

The foraminiferal biofacies of Assemblage B indicates that the basin still had good connections with the open ocean during the latest early to earliest middle Messinian stage II. At the same time, however, the first indications of enclosure of the basin (enlarged fresh-water influence in the surficial waters and development of a salinity gradient) are found. These mark the onset of a fluctuating shift to complete dominance of continental over marine conditions, characterizing late Messinian stage IV.

This development can be compared with central Mediterranean and nearby NW Atlantic sections (Fig. 1).

The rapid increase in waterdepth at the end of the Tortonian (Fig. 7), can be observed in many sections inside the Mediterranean (Van der Zwaan 1982, Glaçon et al. 1990) and outside in the atlantic Guadalquivir and Gharb Basins (Bossio et al. 1977, Glaçon et al. 1990, personal observations).

The earliest Messinian, deep open marine Subassemblage A2, compares with atlantic assemblages of the same age from the Guadalquivir Basin (Verdenius 1970, Berggren & Haq 1976, Glaçon et al. 1990, fig. 3) and the Gharb Basin (unpublished data), and with northern Italy (Colalongo et al 1979) and offshore Lybia (Van Hinte et al. 1980) within the Mediterranean. The Central Sicilian Basin is characterized by the total absence of benthos (Colalongo et al. 1979, Van der Zwaan 1982). Apparently, there was an open connection with the Atlantic, but normal marine conditions existed in the Mediterranean only in the upper few hundred meters of the water-column, whereas anoxity occurred in its deeper parts. Planktonic foraminifer assemblages already show some marked differences with the Atlantic, which have relatively common *Sphaeroidinellopsis* spp. and globoquadrinids and relatively rare *G. conomiozea* (Bizon et al. 1972, Wernli 1977, personal observations), suggesting a separation of the water masses.

The typical buliminacean-rich Assemblage B is characteristic for all Mediterranean sections in the interval encompassing the *N. acostaensis* coiling change (Colalongo et al. 1979, Van Hinte et al. 1980, Orszag-Sperber et al. 1980, Troelstra et al. 1980, Van der Zwaan 1982). At the same time diverse benthic and planktonic assemblages persisted in the Atlantic, where the first primitive G. *margaritae* appear (Wernli 1977, personal observations). The only evidence that the Atlantic had a better connection with SE Spain than with the central Mediterranean is the scarcity of the endemic *T. multiloba* in our area.

The first appearance of G. *margaritae* s.s. outside the Mediterranean approximately coincides with the onset of Mediterranean evaporite deposition (Bossio et al. 1977; Fig. 3.16). Above this level diverse foraminiferal assemblages persist in the Gharb and Guadalquivir basins (Perconig 1966, Verdenius 1970, Berggren & Haq 1976). The Carboneras area had only scarce more open marine conditions during the evaporite deposition. From Italy also a few marine levels have been described from the top of the lower evaporites (Selli 1973, Sturani 1978, Vai & Ricchi Lucci 1978).

The Messinian overlying the evaporites in the basins of SE Spain contains a characteristic biofacies (Sorbas Basin: Roep & Van Harten 1979; Vera Basin: Geerlings et al. 1980 a, b; Nijar Basin: our Assemblage D) comparable to the typical 'lago mare' assemblages described from Italy (Ruggieri & Sprovieri 1976, Colalongo et al. 1978).

In the deposits directly overlying the evaporites in the central part of the Carboneras-Nijar and Sorbas Basins of SE Spain (our Subassemblage D 1), marine fossils are either absent or a few specimens are found of which the autochthonous nature is questionable (Ott d'Estevou, 1980, p. 80/81; this chapter).

The relative abundance of planktonic foraminifers in the youngest Messinian deposits of the Vera and Nijar Basins (Montenat et al. 1976, De la Chapelle 1987) has been demonstrated to be due to reworking (Geerlings et al. 1980 a, b; this chapter).

The absence of marine late Messinian in the deeper parts of the basins, is an important argument to correlate marine intercalations near the top of the marginal deposits of the Agua Amarga, Sorbas and Nijar Basins (Esteban & Giner 198O, Ott d'Estevou 1980, Van de Poel 1991) with marine levels in the 'middle' Messinian Yesares Formation, rather than to consider them late Messinian in age (Montenat et al. 1980).

The basins of southeastern Spain abruptly returned to open marine conditions in the early Pliocene as did the rest of the Mediterranean (e.g. Cita 1973, Casati et al. 1978).

APPENDIX 1: EXPLANATORY NOTE ON THE MICROFOSSIL TAXA

In the distribution charts (App. IIa-c), individual taxa are plotted in larger groups. The following explanation can be given on their composition. Biostratigraphically important taxa are illustrated in Plates I, II and III and their ranges were given in Fig. 4.

Planktonic foraminifers

Globorotaliids are presented as keeled and unkeeled forms in App. IIa-c. The keeled forms mainly belong to the *G. miotumida* Jenkins group (sensu Sierro 1985) with common *G.conomiozea* Kennett (Plate I.13). A few small advanced forms of the *G. menardii* D'Orbigny plexus (cf. *G. menardii* forms 4 and 5 of Tjalsma 1971 and Zachariasse 1975) occur in the lowermost samples, whereas *G. margaritae* Bolli and Bermudez (Plate III.5, 6 and 9) and *G. crassaformis* Galloway and Wissler are present in the upper part of the section (Fig. 4).

Unkeeled globorotaliids are *G. scitula* Brady, the ventrally high *G. suterae* Catalano and Sprovieri, and *G. puncticulata* Deshayes (Plate III.10).

The **neogloboquadrinids** are *N. acostaensis* (Blow) (Plate II.12) with *N. humerosa* (Takayanagi and Saito) as a common companion in the Lower Abad Marl.

Globigerinita spp. commonly occur in samples from the Abad Marl, but have been taken together in App. IIa-c and Fig. 4 with the far more dominant ***Turborotalita quinqueloba*** (Natland) (Plate II.11).

The ***Globigerina*** spp. mainly are *G. bulloides* D'Orbigny (s.l.), *G. obesa* Bolli and *Globigerinella aequilateralis* (Brady), but also include *G. decoraperta* Takayanagy and Saito and *G. nepenthes* Todd. An Acme of the first group is observed in several samples from the lower part of the Upper Abad Marl, where *G. bulloides* often has a bulla.

The ***Globigerinoides*** spp. of the Abad Marl are *G. obliquus* Bolli and *G. obliquus extremus* Bolli and Bermudez with less common *G. trilobus* Reuss (s.l.). In the Pliocene Formations both groups are more evenly distributed. ***Orbulina*** shows a wide range of form variation in the Upper Abad Marl.

Benthic foraminifers

The category "**agglutinants**" contains finely agglutinated species (predominantly *Spiroplectammina carinata* (D'Orbigny), *Martinottiella communis* (D'Orbigny), *Eggerella bradyi* (Cushman) (Plate III.1)), and robust, coarsely agglutinated *Textularia* spp. (Fig. 4).

Nodosariids stands for taxa belonging to the superfamily *Nodosariacea* of Loeblich and Tappan (1964), excluding *Lenticulina* spp.. Fragile forms, with common *Amphicoryna scalaris* (Batsch) and *Dentalina* spp., or robust, heavily costate nodosariids occur in stratigraphically different levels (Fig. 4).

Lenticulina spp.. Large, robust types are restricted to the upper part of the section.

The category '**small deep water taxa**' consists of *Gyroidina neosoldanii* Brotzen, *Melonis* spp. (Plate I.2), *Pullenia bulloides* (D'Orbigny), *P. quinqueloba* (Reuss) (Plate I.3) and *Sphaeroidina bulloides* D'Orbigny.

Cassidulina spp. stands for *C. laevigata* D'Orbigny, *C. subglobosa* Brady and *C. crassa* D'Orbigny (in descending order of abundance). Besides, few *C. oblonga* Reuss have been found.

Representatives of the genus **Cibicidoides** generally occur either as *C. pseudoungerianus* (Cushman) (Plate I. 9), or as *C. dutemplei* (D'Orbigny) (Plate III. 2) (Fig. 4). In Po 153 and PK 81 of Section 1 both types are common. Intergradation, as earlier suggested by Van der Zwaan (1982), is here observed.

As **'epiphytes'** have been counted: the robust *Cibicides lobatulus* (Walker and Jacob) (Plate VI, Fig. 8) and the smaller taxa *Hanzawaia boueana* (D'Orbigny) (Plate II.3), *Asterigerina planorbis* D'Orbigny (Plate II.15) and *Rosalina globularis* D'Orbigny.

The few miliolids (mainly *Quinqueloculina* sp.) and *Cancris auriculus* (Fichtel and Moll) and *Valvulineria bradyana* Brotzen, which show the same distributional pattern as the 'small epiphytes' in the Upper Abad Marl, have been included.

The ***Uvigerina flintii*** Cushman group comprises the large, inflated uvigerinids with fine striation and a glossy appearance. Those from the lower part of the section are *U. striatissima* Perconig (Plate I.5), whereas those of the Pliocene pertain to *U. longistriata* Perconig (Plate I.4).

Uvigerina peregrina includes completely costate forms (Plate I.11) and forms which lack costae on the last chamber (Plate I.12a, b). The latter may pertain to *U. peregrina* Cushman var. *dirupta* Todd, *U. hispido-costata* Cushman and Todd (cf. Poore and Stone 1981) or still another variety of the *U. peregrina* plexus (cf. *Uvigerina* sp. 221 of Lutze 1986). In Assemblage B from the Upper Abad Marl small specimens and forms with poorly developed costae are encountered (Plate II.1a, b), which are recorded as *Uvigerina* cf. *peregrina* in App. IIa-c. Forms with poorly developed costae have also been found on shallow levels of the Peru-Chili Trench (*Uvigerina peregrina* var. A of Ingle et al. 1980).

Bulimina aculeata D'Orbigny contains both, the more typical form (Plate II. 2) and *B. aculeata* D'Orbigny var. *minima* Tedeschi and Zanmatti (Plate II.8).

Bulimina elongata D'Orbigny (Plate II.7) shows intergradation with *B. aculeata* (Van der Zwaan 1982) and it is here suggested that it represents an ecophenotypic variant. It may be characteristic for the shallow, slightly hyposaline waters from which it is consistently reported (Murray 1973, Pujos 1976, Haake 1977, Moodley 1990)

The ***Bolivina spathulata*** **group** consists of the short, robust, *B. dilatata* Reuss (Plate I.7) and, less common, *B. subreticulata* Parr, as well as the more elongate and, especially, thinner, *B. spathulata* (Williamson) (Plate II.9) and the short, thin, *B. dentellata* Tavani (Plate II.10).

Elphidium spp. contains the small species *E. fichtellianum* (D'Orbigny) (Plate I.4), *E. aculeatum* (D'Orbigny) (Plate II.4) and *Elphidium* sp. A (Plate II.17), as well as the robust *E. crispum* (Linnaeus).

Elphidium sp. A resembles a Recent Elphidium (Plate II.16) which is common in pre-concentration basins of natural salt works of the Mediterranean coast of southern France (Zaninetti 1983, personal observations in the "Salin de Gruissan", near Narbonne). The latter has been desribed as *E.* cf. *alvarezianum* D'Orbigny by Zaninetti (1973). It should be remarked that similar morphotypes have been described as *E. margaritaceum* Cushman, *E. articulatum* D'Orbigny and *Cribrononion* cf. *alvarezianum* from a coastal section of the Adriatic and diverse hyposaline lagoons and estuaries (Jorissen 1988, Murray 1973, Lutze 1965).

Ammonia spp. stands for *Ammonia beccarii* (Linnaeus) and *A. tepida* Cushman (Plate II.18).

Ostracoda

Two groups have been recognized: the genus *Cyprideis* and a group 'Other'. Within the latter, shallow marine *Aurila* sp. are chracteristic for turbiditic levels of the Upper Abad Marl, whereas the late Messinian brackish 'lago mare' forms *Loxoconcha djaffarovi* Schneider and *Tyrrhenocythere pontica* Liventhal (Ruggieri & Sprovieri 1976; Carbonnel in Montenat et al. 1976, Roep & Van Harten 1979), are present at the top of the Feos Formation.

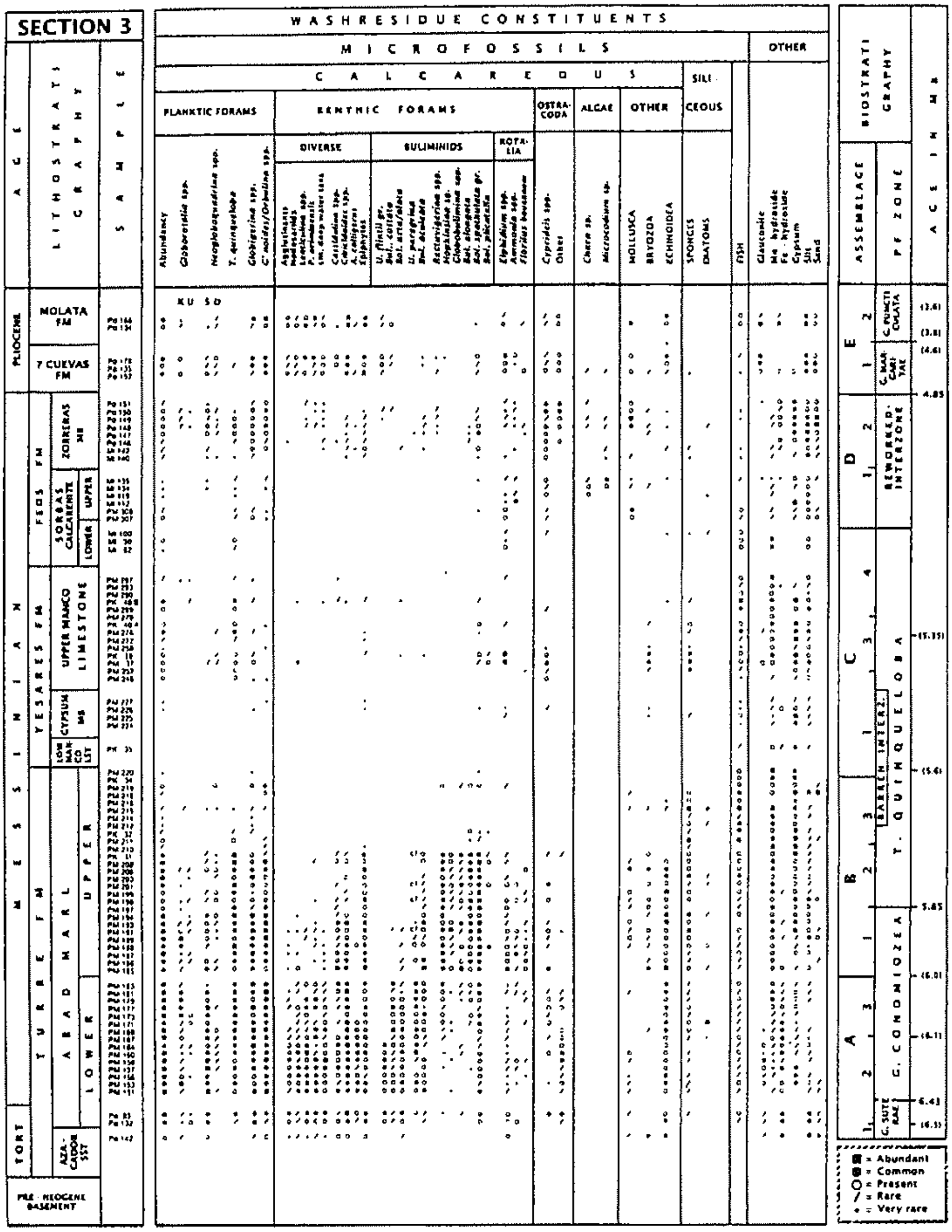
Algae

'**Algae**' stands for calcite fragments that are the product of the continental bacterium *Microcodium*, and for complete or fragmented oogonia of the fresh- to brackish water plant *Chara* sp..

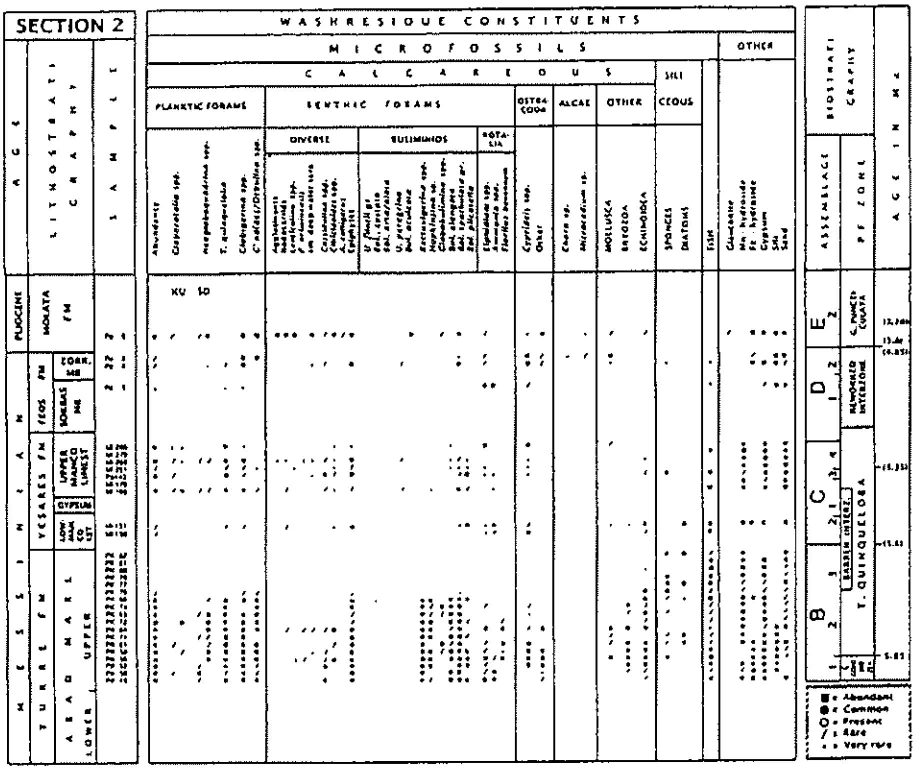
Other

Additional fossil elements of the washed residues are fragments of **calcareous macrofossils**, remains of **siliceous organisms** and of **fish**. The pink, thin-shelled gastropods and bivalves from the Feos Formation are remains of typical 'lago-mare' mollusks (Ruggieri & Sprovieri 1976, Colalongo et al. 1977). 'Micromulluscs' with paratethyan affinity have been earlier reported from the top of the Messinian of the adjacent Sorbas Basin (Iaccarino et al. 1975, Archambault Guezou 1976, Ott d'Estevou 1980).

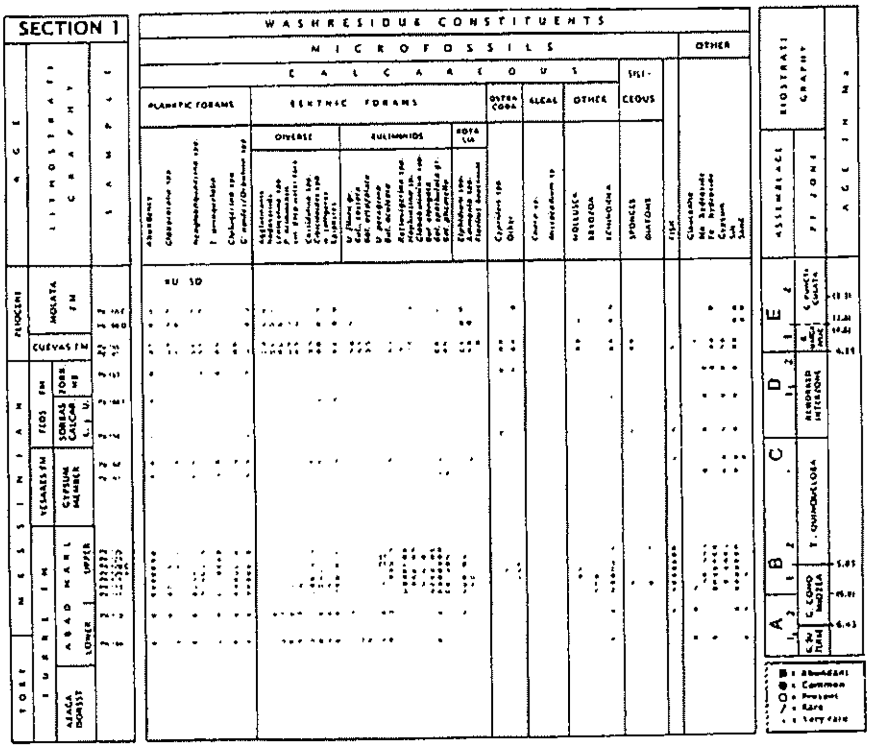
APPENDIX II: MICROFOSSIL DISTRIBUTION CHARTS



App. IIa. Microfossil distribution chart and zonation of Section 3. Under *Globorotalia* K=keeled, U=unkeeled; under *Neogloboquadrina* S=sinistral, D=dextral.



App. IIb. Microfossil distribution chart and zonation of Section 2. Under *Globorotalia* K=keeled, U=unkeeled; under *Neogloboquadrina* S=sinistral, D=dextral.



App. IIc. Microfossil distribution chart and zonation of Section 1. Under *Globorotalia* K=keeled, U=unkeeled; under *Neogloboquadrina* S=sinistral, D=dextral.

3.4. SEQUENCES AND PARASEQUENCES IN THE MIO-PLIOCENE OF THE NIJAR BASIN (S.E. SPAIN). EUSTATICS AND TECTONICS IN A FORMER SEAWAYS CONNECTING MEDITERRANEAN AND ATLANTIC[1]

*H.M. van de Poel**

INTRODUCTION

The Nijar Basin is located at the far end of the former seaways connecting Atlantic and Mediterranean. The Mio-Pliocene transition takes place in a tectonically quiet period. At the beginning of the Messinian, the area shows little topographic relief and is largely covered by a shallow sea.

DESCRIPTION

The following sequences and depositional units can be recognized and interpreted in terms of sequence stratigraphy (from old to young, Fig. 1):

------- 'sequence 0' (Tortonian), is only fragmentically exposed. In the neighbouring Tabernas Basin a submarine fan is well-developed. The youngest deposits are marine hemipelagic marls with some turbidites, that probably make out part of the transgressive systems tract, truncated by a regional unconformity.

------- Sequence A; Late Tortonian thick-bedded graded sandstones with shallow marine fossils represent the lowstand facies. They are overlain, progressively unconform towards the basin margin, by marine marls, that are laterally replaced by a transgressive series of basal conglomerates and sandstones which truncate older Neogene sediments in more central parts of the basin and cover an irregular surface of pre-Neogene basement versus the margin. this represents the transgressive systems tract. Overlying Messinian marine hemipelagic marl and a laterally equivalent regressive outbuilding reef complex, in which a number of parasequences occur, can be interpreted as the highstand systems tract. It should be noted that this interval shows a relatively strong regressive tendency. This is expressed both in the regional downstepping of reefs to a ca. 100 m lower topographic level and a strong upward shallowing in the basinal marls indicated by a drastic change in composition of the benthic foraminiferal faunas. The oligotypical character of both the latter and the reef fauna attest of the installation of a restricted basin during this regression.

------- Sequence B (middle Messinian age); Lowstand facies is represented by an alternation of laminated marls, that may contain a very shallow marine microfauna, with thick gypsum beds. The transgressive systems tract is formed by laminated marls with a relatively deeper restricted marine fauna and thinner gypsum intercalations and laterally equivalent oolitic and some bioclastic limestones, locally with small coral reefs, that overlies the older reef complex with an erosional unconformity. A number of parasequences are here again developed. The top of this sequence shows regressive outbuilding of calcarenites and shallow marine microfossils.

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[1] Text and figure earlier published in *Strata, Série 1, Vol. 5(1989), pp. 153-155.*

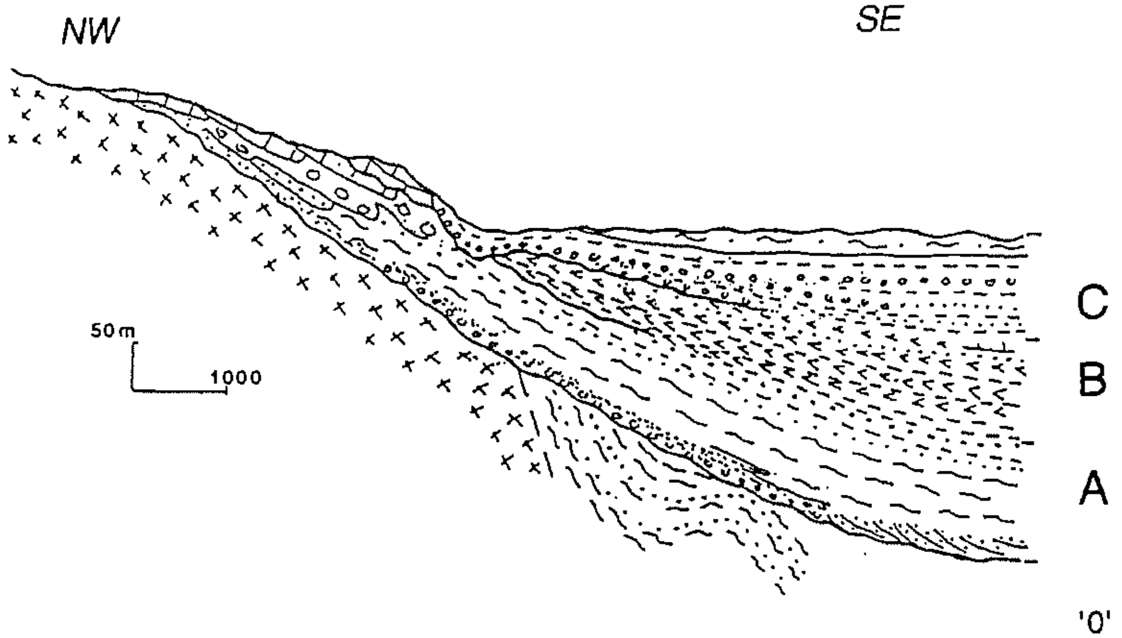

Fig. 1. Simplified geologic cross-section through the central and marginal part of the Nijar Basin, showing its Late Neogene sedimentary fill and its subdivision in depositional sequences

—— Sequence C starts with lake deposits with river and deltaic intercalations. It is comparable to a lowstand fan or delta system but deposited entirely under continental conditions. An important unconformity underlies these deposits towards the basin margins. A conformable cover of earliest Pliocene silty marl with planktonic foraminifera, indicating waterdepth of some hundred meters in the base, represents the transgressive system tract. It is truncated by an important regional unconformity.

CONCLUSIONS

The commonly strong vertical and horizontal shifts in depositional depth at the boundary of depositional units and sequences and the often truncated aspect of the latter are sometimes the effect of tectonics (truncation below, and rapid transgression in, the base of sequence A; truncation at the top of sequence C). Other boundaries are interpreted as the effect of important sealevel movements within the Mediterranean, after isolation or reopening initiated by eustatics (sequence boundary at the base of C, probably correlating with the base of TB 3.3; rapid transgressions within sequences B and C). The nature of the boundary between sequence A and B suggests a relatively gradual event. Both the common occurrence of parasequences and the great difference in facies between sequences B (dry evaporitic) and C (humid conditions) attest to climatic variations, probably related to a glaciation event.

CHAPTER 4

THE SORBAS, VERA & AGUA AMARGA BASINS

4.1. INTRODUCTION

The Sorbas and Vera basins contain the type-sections of most of the 'classical' lithostratigraphic units of the area (Ruegg 1964, Völk & Rondeel 1964, Rondeel 1965, Völk 1967). In particular the Sorbas Basin has a well-exposed record of the Messinian, whereas also good sections of 'older Miocene' and earliest Pliocene sediments can be found in these basins (Völk 1967, Montenat 1973a, *et al.* 1976, 1980 Iaccarino et al. 1975, Dronkert 1978, 1985, Ott d'Estevou 1980, etc.). Differences of opinion still exist on the degree of isolation of these basins during the Messinian, and, especially, as to the relative importance, origin and age of unconformities in its record (op. cit. and Geerlings et al. 1980 a & b; Müller & Hsü 1987; Barragan et al. 1990, Ott D'Estevou et al. 1990, Benson et al. 1991, Riding et al. 1991; Fig. 1.4).

The study of these basins also allows for the evaluation of the difference in Messinian facies development between a more internal basin of the Betic Cordilleras (Sorbas Basin), as compared with the coastal Vera Basin and of the development of the E-W trending Miocene marine corridors connecting the Mediterranean with the Atlantic through the internal part of the Betic Cordilleras (Fig. 2.1).

The Agua Amarga Basin is well-suited for basin analysis because of its high degree of exposure and the relatively small tectonisation of the basin margins, whereas also Messinian diagenetic features are well-developed. However, no comprehensive, detailed study of this basin was published. Bousquet et al. (1975a), Bousquet & Philip (1976), Montenat et al. (1976), Roux & Montenat (1977), Addicott et al. (1978), Dabrio & Martin (1978), Dronkert et al. (1979), Estaban (1980), Armstrong et al. (1980), Pineda Velasco et al. (1983), Dronkert (1985), De la Chapelle (1988), Montenat & Ott D'Estevou (1990) and Riding et al. (1991) have discussed certain aspects of its Neogene sediments as part of studies with a wider framework. Facies and stratigraphy of the Mio-Pliocene section from more central parts of the basin, with emphasis on the Mio-Pliocene boundary interval, was discussed by Van de Poel (1980; in Pineda Velasco et al. 1983) and Van de Poel et al. (1984). Esteban & Giner (1980) and Franseen & Mankiewicz (1991) have given detailed descriptions of Messinian carbonate margin deposits.

A comparison of the stratigraphy of the Agua Amarga Basin with that of the Northern Nijar Basin offers a further opportunity to evaluate the difference between a more distal and proximal basin in respect to the Mediterranean.

In the following five subchapters, we first compare the Messinian stratigraphic successions of these basins, with emphasis on the evaporitic series, as well as with the type-Messinian (Dronkert et al. 1979; Ch. 4.2). We underline the abundance of typical 'Mediterranean' Messinian facies, with common cyclic development, in the inland Sorbas and Northern Nijar basins and their relative paucity in the coastal Vera and Agua Amarga basins and discuss biostratigraphic correlation between the Sorbas and Nijar basins on the one hand and the Vera Basin on the other, showing the possibility of an important intra-Messinian hiatus in the latter. Some discussion is also given of the relative importance of different underlying factors (eustacy, global climatic changes, local tectonics and Mediterranean sealevel lowering) in the facies development of the EAP and the implications of the latter for the interpretation of the "scenario" of the Messinian Salinity Crisis in the Mediterranean.

In a second subchapter, the early Messinian facies development of the **Sorbas Basin**, with emphasis on the foraminifer composition in the central part of this basin is described (Troelstra et al. 1980; Ch. 4.3), and an interpretation of the depositional environment is given in terms of waterdepth, salinity and temperature. The paleogeography of the sill area and the relative importance of the local sedimentary factor on the development of sill-depth in the Entrance Area are discussed.

A first biostratigraphic framework for the Mio-Pliocene of the EAP, is given in Geerlings and Van de Poel 1979 (Ch. 4.4.1). Its application in the Mio-Pliocene boundary interval in the **Vera Basin** further underlines the particular development in this area, where either an important hiatus must be present or extremely condensed sedimentation in a non-marine environment must have taken place during the middle-late Messinian. New field data presented in Chapter 4.4.2 (Geerlings et al. 1980) confirm the first hypothesis and strongly suggest the importance of Mediterranean sealevel lowering on the stratigraphic development in this area.

The Mio-Pliocene stratigraphy of more central parts of the **Agua Amarga Basin** is described in Chapter 4.5 (Van de Poel et al. 1984) with special attention to the presence of a remarkable limestone breccia, soil horizons and important river-incision at the Mio-Pliocene boundary. Arguments for limestone brecciation by (evaporite-)dissolution-collapse are presented and we discuss the relative importance of local tectonics, regional humid climate and Mediterranean sealevel lowering in the facies development.

4.2. GYPSUM DEPOSITS IN THE PROVINCE OF ALMERIA; CONSEQUENCES FOR THE WESTERN MEDITERRANEAN[1]

H. Dronkert, H.M. van de Poel & L.P.A. Geerlings.

SUMMARY

A brief survey of the facies development of the Late Miocene-Early Pliocene sedimentary sequences is given for two inland basins, the Sorbas Basin and the Campo de Nijar area, and two basins bordering the Mediterranean coast, the Vera Basin and the Agua Amarga Basin. The inland basins display a typical 'Messinian' sedimentary facies in which a strong cyclic development is observed. The coastal basins generally lack such sedimentary facies and cyclic developments, especially the Vera Basin, for which the possibility of a hiatus is discussed.

A 'continuous inflow' model, allowing for a casual dessication, is proposed for the observed development and facies of the evaporites in these 'Margin basins'. The 'dessication only' model poses problems for the explanation of the facies observed in the evaporites of the shallow Margin basins in SE Spain.

INTRODUCTION

In SE Spain the Permo-Triassic and the Messinian represent periods with intense evaporite development. In other periods in this area evaporites are of minor importance or have all vanished due to erosion. The Permo-Triassic evaporites are severely tectonized in shear and fracture zones or intensely metamorphosed, which makes detailed genetic studies difficult. The Messinian evaporites are known from the Mediterranean region and from areas far into Asia. In many of these areas these evaporites are strongly tectonized. The deeper buried evaporites have also been partly or wholly recrystallized (metamorphosed). Thus the deep-lying evaporites (drillcore samples) and samples from tectonized areas create problems for evaporite genetics and correlation. In some areas, amongst which the Sorbas Basin, those problems are lacking or can be easily overcome. The facies of the gypsum in the Sorbas Basin as well as the basin geometry indicate a subaqueous depositional environment of about 100 m waterdepth (Dronkert 1976, 1977, Pagnier 1976).

If one considers the West Mediterranean as one basin in which evaporation has taken place simultaneously in the deeper part and in the coastal areas ('Margin basins'), as accepted by Hsü et al. (1973a), it seems impossible to explain the present distribution of the various evaporites. Before precipitation of gypsum begins, the water level in a totally isolated West Mediterranean basin will have dropped below the bottom of the Margin areas, in which now Late Miocene gypsum is found (these basins belonged to the rim of the bull's eye pattern of Hsü et al. 1973a). Generally the gypsum in the Margin basins in SE Spain precipitated as primary gypsum in less than a few hundred meters of marine water (Montenat 1973a, Dronkert 1976, 1977, Schleich 1977). This is in contrast with a secundary recrystallization of the gypsum or an origin of the gypsum in 'hanging' basins fed by continental waters, as would be implied by the 'dessication only' model for the evaporites

[1]Text earlier published in *Ann. Géol. Pays Hellén., Tome hors série (1979), fasc. 1, pp. 345-354.*

in the Margin basins. Thus the evaporites of the Margin basin and the deep-water evaporites were not deposited as a result of one dessication phase. THIS IMPLIES A SEPARATE DEVELOPMENT OF THE EVAPORITES AT THE MARGIN AND IN THE DEEP BASINS. The Margin evaporites should then have developed before or after any 'total dessication' of the West Mediterranean basin.

Another possibility is to abandon the 'dessication only' model and work with a continuous inflow model, with or without reflux. A Betic, Rif, or Gibraltar sill would then have regulated the influx of water from the Atlantic Ocean, mixing it with that of the Mediterranean. A similar model has already been proposed by Schmalz (1969), Selli (1973), and for the 'gessi inferiori' by Ruggieri (1967), Hsü et al. (1977) and Van Couvering et al. (1976, first proposed and later rejected). A shallow sill would make fluctuations of the water level in the Atlantic Ocean of extreme importance for the water circulation in the Mediterranean Sea. Freezing and melting of polar icecaps possibly caused by solar activity, could then have regulated the water supply to the Mediterranean basin, and have generated a cyclic development that will have to be visible in the sediments of that area.

And indeed, a strong cyclic development is observed in most of the Mediterranean evaporites (e.g. see in Drooger 1973 and Catalano et al. 1976). The solar dependancy of this model implies no restriction upon the number of these cycles. Local factors, as water circulation, tectonics, sediment provenance and climate place constraints on the character of these cycles and the numbers present. An accidental coincidence of the number of cycles from different areas tells more about the rate of basin development than about time-correlation.

FACIES DEVELOPMENT OF THE MIO-PLIOCENE IN THE SORBAS BASIN AND THE CAMPO DE NIJAR AREA

Facies of the pre-Gypsum deposits

The marls directly underlying the gypsum series in the centre of the Sorbas Basin and in the NE Campo de Nijar, bear a marked resemblance to deposits underlying 'Messinian' evaporites from many other Mediterranean areas. They are characterized by the presence of diatomitic intercalations (Van de Poel 1976, Geerlings 1977, Burckle 1977a), that are also known from other adjacent areas (Campo de Almeria: Iaccarino et al. 1975 and Agua Amarga Basin). Microfauna assemblages, both benthonic and planktonic, are oligotypical (Van de Poel 1976, Geerlings 1977, Civis et al. 1979a). An elevated salinity was deduced from the benthic foraminifer assemblage by Troelstra & Van Gorsel (1977).

Esteban et al. (1977), describing reef deposits of the basin borders, proposed a fluctuating salinity to explain the oligotypical fauna assemblages. We consider these reefs coeval with the latest pre-gypsum sediments of the basin centre. A similar fluctuation of the salinity level could be responsable for the extreme variability of P/B ratios as noted by van de Poel (1976) and Geerlings (1977), and could also be the cause of fluctuations in benthic foraminifer assemblages (Troelstra & Van Gorsel 1977) and the presence of some sterile layers. The cyclic outbuilding of reefs as noted by Esteban et al. (1977) from some areas (e.g. Nijar) could very well reflect regressive and transgressive trends.

Facies of the Gypsum deposits.

The facies development and cyclicity of the gypsum deposits in the Sorbas Basin, the Campo de Nijar area, as well as other adjacent areas as the Andarax Basin and the Campo de Almeria, are very similar to many other Messinian evaporites of the Mediterranean area.

In the central part of the NE Campo de Nijar, evaporitic limestones with pseudomorphs after gypsum and collapse breccias are closely related to the gypsum deposits (Mallee & van de Poel 1974, Dronkert 1976, Van de Poel 1976,, Mallee 1977). These sediments resemble the typical 'calcare di base' described from the Sicilian Messinian.

Facies of the post-Gypsum deposits.

The sediments directly overlying the gypsum sequence in the central part of the Sorbas and Campo de Nijar basins still reflect a high salinity. Cyclic development is also present in transgressive and regressive sequences of these shallow marine deposits (Pagnier 1976, Dronkert 1976, Van de Poel 1976, Roep & Beets 1977, Mallee 1977).

Towards the top in Sorbas, and more laterally in the Campo de Nijar, these sediments are replaced by continental deposits (Roep & Beets 1977, Pagnier 1976, Van de Poel 1976, Mallee 1977) and some lagoonal intercalations. Ostracodes from the last mentioned levels indicate a brackish or still hypersaline environment (cf. 'lago mare').

After deposition of the latter sediments, a return to a normal marine environment is observed in calcareous sandstones, silts and conglomerates, rich in normal marine micro- and macro-fauna, that are generally ascribed to the Pliocene (Van de Poel 1976, Dronkert 1976, Montenat & Ott D'Estevou 1977).

FACIES DEVELOPMENT AT THE MIO-PLIOCENE BOUNDARY IN THE VERA AND AGUA AMARGA BASINS

In the Vera and Agua Amarga basins, that border the Mediterranean, the typical 'Messinian' sedimentary sequence is far less developed:

In the Agua Amarga Basin (south of Carboneras) relatively thin ('calcare di base' like) limestone intercalations are found in the top of locally diatomaceous Messinian marls. Gypsum deposits were not found in this basin. A coarsening upward sequence of sandstones and conglomerates covers reef limestones in the centre of the basin and the locally present, possibly evaporitic limestones mentioned above. It bears resemblance to the coarsening upward sequences in central parts of the Campo de Nijar that overlie the gypsum and evaporitic limestone series in that area. Pliocene fossiliferous calcareous sandstones and conglomerates are the youngest marine sediments in this basin.

In the Vera Basin, typical Messinian evaporites are very scarce. Gypsum remains are only present along the southern Mio-Pliocene boundary, but their stratigraphic position is still not fully understood.

A sandstone and conglomerate sequence decribed by Völk (1967) has the same lithological aspects as the coarsening upward sequence described from the Agua Amarga Basin. It covers regressively Messinian reef deposits of the western basin border and is overlain by Völk's early Pliocene 'Cuevas Formation'.

'Lago mare' deposits are described from the latest Miocene of the Cuevas del Almanzora section in the central part of the basin (Cita et al. 1978). These deposits are however still highly controversial (Montenat et al. 1976). Marly and sandy sediments overlying the gypsum remains of the Coscojar area (see Dronkert 1976) contain a microfauna resembling in some aspects that of the uppermost Miocene of the Cuevas del Almanzora section (*Cyprideis* fauna, accompanied by *G. conomiozea*).

AGE OF THE 'MESSINIAN' EVAPORITES

A terminal Miocene age is generally assigned to the gypsum series of the Sorbas Basin and adjacent areas (Montenat 1973, Iaccarino et al. 1975, Dronkert 1976, Pagnier 1976, ADARO maps) and to the marls directly underlying the evaporites. These marls resemble closely those underlying Messinian evaporites of Italy, both in facies and fauna, of which the micro-plankton assemblage was compared to that of the *G. multiloba* subzone of D'Onofrio et al. (1975) (Geerlings 1977, Gonzalez Donoso & Serrano 1977, Civis et al. 1979a).
However, a local correlation of planktonic foraminiferal assemblages still gives rise to somewhat controversial results in our opinion.

The change in coiling direction of *Gl. acostaensis* (= *G. dubia* Hermes non Egger) from preferentially sinistral to dextral was observed in the upper part of the marls directly underlying the gypsum in the Sorbas Basin (Hermes in Völk 1967, Geerlings 1977, Gonzalez Donoso & Serrano 1977). Völk (1967) noticed the same phenomenon at the boundary between his youngest Miocene 'Turre Formation' and his oldest Pliocene 'Cuevas Formation'. It is also described from the adjacent Campo de Nijar by Civis et al. (1979a) as well as from other areas in southern Spain (e.g. Lorca: Manuputty 1977).

From Italy it is described from what would seem to be the same stratigraphic level, in the base of the 'Tripoli' of the Capodarso section of Sicily (Zachariasse 1975), while Bossio et al. (1977a) describe this change in coiling direction from the early Messinian of the Piedmont basin of N. Italy.

Its stratigraphic value on a wider scale is (amongst others) further stressed by Bossio et al. (1977a), who describe this change in coiling direction from the Atlantic Moroccan Bou Regreg section. Cita (1977) uses this change of coiling direction as means of correlation of a part of DSDP Site 397 (off Cap Bojador), off W. Sahara, with the Italian 'Messinian'. It seems even correlatable to Pacific pison core sections, where this change of coiling direction is observed in sediments ascribed to the base of magnetic epoch 5 (Cita 1977).

In the Vera Basin, however, this change of coiling direction (see Bizon et al. 1975, Gonzalez Donoso & Serrano 1978), takes place at about what is considered by most authors (e.g. Bizon & Bizon in Montenat et al. 1976, Cita et al. 1978, excluding Perconig et al. 1977) to be the Mio-Pliocene boundary. Völk (1967) and Gonzalez Donoso & Serrano (1978) find it hard to pinpoint the Mio-Pliocene boundary, that they consider as poorly biostratigraphically designed, precisely in this section.

A correlation between the Sorbas and Vera basins on basis of this change in coiling direction leads to three possibilities:

a. Assuming continuous sedimentation in both sections (as has been assumed by Völk 1967 and Bizon et al. 1975 for the centre of the Vera Basin), the change in coiling direction of *Gl. acostaensis* as found in the marls directly underlying the gypsum in the Sorbas basin would correlate to that found at the base of the 'Cuevas Formation', which is considered as lower Pliocene in age by most authors. The gypsum of the Sorbas Basin would then also be of Pliocene age. Another possibility to explain this apparent contradiction is to deny the stratigraphic value of this change in coiling direction as it is not even possible to correlate two adjacent basins in SE Spain on basis of this phenomenon.

b. Assuming an hiatus in the Vera Basin between 'Turre' and 'Cuevas' Formations, the level at which the change of coiling direction of Gl. acostaensis is observed in the upper part of the marls underlying the gypsum in the Sorbas Basin (and also the same level in the lower part of the Messinian type-section and in all the sections described above), would then correlate to an 'apparent' change in coiling direction at the boundary between the 'Turre' and 'Cuevas' Formations.

The upper part of the 'Turre Formation' would then be older than the upper part of the marls underlying the gypsum in the Sorbas Basin and of the 'Tripoli Formation' of the Messinian stratotype. The gypsum of the Sorbas Basin could then well be of Messinian age.

c. Assuming an hiatus between a level somewhat below the top of the 'Turre Formation' in the Vera Basin and its top consisting of 'lago mare' deposits as assumed by Cita et al. (1978), would correlate the level at which the change in coiling direction is observed in the upper part of the marls underlying the gypsum in the Sorbas Basin, and also the same level in all the sections described above, including that in the lower part of the Messinian stratotype, to the level somewhat below the top of the 'Turre Formation'.

The microfauna from the top of the 'Turre Formation' would then have to be considered reworked, as stated by Cita et al. (1978). The lower part of the 'Turre Formation', below the 'lago mare' deposits, would then also be older than the top of the marls directly underlying the gypsum in the Sorbas Basin and of the 'Tripoli Formation' of the Messinian stratotype. In this case a 'Messinian' age is also highly likely for the evaporites of the Sorbas Basin and adjacent areas.

DISCUSSION

As described above, the continuous inflow model seems to offer the best explanation for the genesis of the evaporites in the SE Spanish basins. However, there seems to be room for a 'total dessication' during latest Miocene times.

If the evaporites of the Margin basins developed after a 'total dessication', evidence of this dramatic event should be visible in the pre-evaporitic sediments of these basins. No such evidence has been found.

If the evaporites in the Margin basins developed immmediately before a 'total dessication', the effects of such an event should be visible within the evaporites of these Marginal basins (or there would not be any evaporite, as suggested by Van Couvering et al. 1976).

The effects of a 'total dessication' of the West Mediterranean Basin on the 'Marginal' basins would at least result in karstification of reef limestones and gypsum, in erosional surfaces or in a sedimentary hiatus. However, it is very difficult to decide whether these phenomena are the result of a 'total dessication' or due to local causes (e.g. tectonics). Dissolution of reefs, karstification of gypsum, as well as erosional surfaces and relicts of vanished evaporites are all found in the Alicante area (Esteban et al. 1977, Montenat 1973a, Orti Cabo 1977), in the Vera Basin (Montenat et al. 1976, Vôlk 1967, Dronkert 1976, Cita et al. 1978) and in the Sorbas Basin (Dronkert 1977b, Montenat & Ott D'Estevou 1977).

Some of these features however still lead to controversial interpretation.

4.3. PALEOECOLOGICAL CHANGES IN THE LATEST MIOCENE OF THE SORBAS BASIN, SE. SPAIN[1]

S.R. Troelstra, H.M. van de Poel, C.H.A. Huisman, L.P.A. Geerlings & H. Dronkert

RESUME

Au 3e Séminaire sur le Messinien (Malga, 1977), Geerlings a présenté la distribution des foraminifères planctoniques dans la coupe de Los Perales située dans le bassin de Sorbas. Une étude quantitative de la microfaune benthique de cette coupe a été réalisée maintenant pour servir de standard pour nos recherches dans les bassins voisins. Les résultats préliminaires de l'analyse paléoécologique et leurs relations avec la lithologie sont discutés.

Plusieurs explications peuvent être proposées pour les changements lithologiques et faunistiques intervenant à la fin du Miocène: variations de salinité, de profondeur ou de température; ces trois facteurs ont vraisemblablement joué un role effectif; un accroisement de la salinité dans un bassin dont la profondeur diminue, peut-être accompagné d'un refroidissement, est intervenu pendant le dépôt de la partie supérieure de l'intervalle considéré. Des recherches préliminaires dans les bassins voisins (Vera, Campo de Nijar) semblent indiquer que ces phénomènes ont une extension géographique plus importante.

SUMMARY

In a previous presentation at the Messinian Seminar 3 (Malaga, 1977), Geerlings described the distribution of planktonic foraminifera in the Los Perales section of the Sorbas Basin. The same section has now been analysed quantitatively for its benthonic foraminifera content, to serve as a standard for our investigation in the adjoining basins. The preliminary results of a paleoecological analysis are discussed in relation to the lithological facies development.

Various explanations can be offered for the faunal and lithological changes in the latest Miocene: change in salinity, change in depth, change in temperature. It seems most likely, that all three factors mentioned were effective; an increased salinity in a shalloing basin, possibly accompanied by a pronounced cooling, set on during deposition of the upper part of the studied interval. Initial investigations in neighbouring basins (Vera, Campo de Nijar) seem to indicate that comparable trends are present over a wider area.

INTRODUCTION

At the occasion of the Messinian Seminar III (Malaga 1977), Geerlings described the distribution of planktonic foraminifers in the Los Perales section in the centre of the Sorbas Basin. Towards the top of the section - above the lowermost Messinian (sensu Colalongo et al. 1979) - a conspicuous change in paleoenvironment can be deduced from the fauna development. A roughly contemporaneous change in paleoenvironment was described from other parts of the Sorbas Basin, as well as from its adjoining areas (e.g. Montenat 1973a, Iaccarino et al. 1975, Perconig 1976, Rouchy 1976, Dronkert 1976, Dronkert et al. 1979, and

[1]Text and figures earlier published in *Géologie méditerranéenne 7 (1980), pp. 115-126.*

various papers of Messinian Seminars I-V and the VIIth Int. Congress on the Mediterranean Neogene, of Athens, 1979). However, opinions of various authors on the relative importance of this change, its causes and its implications for the paleogeographic connections between the Mediterranean and the Atlantic (Gibraltar/Betico-Riffian straits) still differ widely.

In order to obtain a better understanding of the sequence of events during the early Messinian and their causes, a quantitative study of the benthonic foraminifera of the Los Perales section was made. In addition, a similar study was undertaken on samples from the Hueli section in the SW of the Sorbas Basin, to see whether the faunal changes observed in the Los Perales section had a more regional character and to see whether any difference could be detected in the development of facies in more central (Los Perales) versus more marginal (Hueli) parts of the basin. Some new data on the planktonic foraminifera from the Los Perales section are also presented.

In this chapter the paleoecological analyses are concentrated on three main factors: salinity, temperature and depth of deposition. The preliminary results will be discussed in relation to the lithofacies development of the Sorbas Basin and that of the adjacent basins.

A compilation of the stratigraphy of the area, largely based on the work of above-mentioned authors and our own findings, will also be given.

MATERIALS AND METHODS

During the course of the study, 48 samples of the Los Perales Section and 33 from the Hueli Section have been analyzed for their foraminifer-content. Washing of the samples followed standard procedures. The remaining washresidue was split into fractions, until 200-300 specimens were left. All these specimens were identified and counted, after which percentages for each species were calculated. Most data on the planktonic foraminifers are from Geerlings (1977).

STRATIGRAPHY

Both sections (Fig. 1) were taken in a predominantly hemipelagic marl sequence. This sequence was deposited in the more central part of the Sorbas Basin, before the onset of Messinian evaporite deposition. Several samples from the lower part of the sections contain forms of the *G. conomiozea* group and this interval can thus be correlated to the lowermost part of the Messinian of Italy. In the lowermost sample of the Los Perales Section such forms are absent (Geerlings 1977). This interval is possibly correlatable to the uppermost Tortonian. The upper parts of the section contain a fauna comparable to that described by various authors from fine-grained sediments directly underlying the evaporite deposits of the Mediterranean area (see Geerlings 1977). A change in coiling direction of the *N. acostaensis* group - from sinsistral to dextral - is present in the top of both sections; A similar change is reported from the Tripoli Fm. in Sicily (Zachariasse 1975, Colalongo et al. 1979).

The base of the sections consists of more or less sandy, bioclastic limestones, deposited during a transgressive phase. This phase followed the late Tortonian tectonization and, at least partial denudation of older Neogene rocks of the Sorbas Basin and adjoining areas (e.g. Vera and Campo de Nijar basins). The marly deposits deposited shortly after, and in part contemporaneous with the transgression, pass laterally into bioclastic limestones, that were deposited in the shallow parts of the basin. Towards the end of the early Messinian, a shallowing tendency is observed at the border of several basins. This is reflected by reef complexes outbuilding towards the centre (Sorbas, Campo de Nijar, Vera, Agua Amarga and Almeria basins). At about the same time, a change in salinity is notable. This is deduced from the oligotypical reef facies at the basin borders and from the transition of more or less normal marine marls into evaporitic deposits in the centre of the basins. Fluctuations in salinity are reflected by the alternation of sulfatic and carbonatic evaporites. Normal marine faunas present in the last-mentioned sequence are generally reworked. However, the presence of some autochthonous faunas, reflecting temporary restoration of more or less normal marine conditions cannot be excluded. Eventually, also the sequences in the central parts of the basins of the area show a shallowing tendency, as they are terminating in coastal , lacustrine and fluvial or other continental deposits overlying the evaporite sequences. At least locally, hiatuses are present in the above-described sequences, but no complete agreement exists on their importance and the causes of their presence.

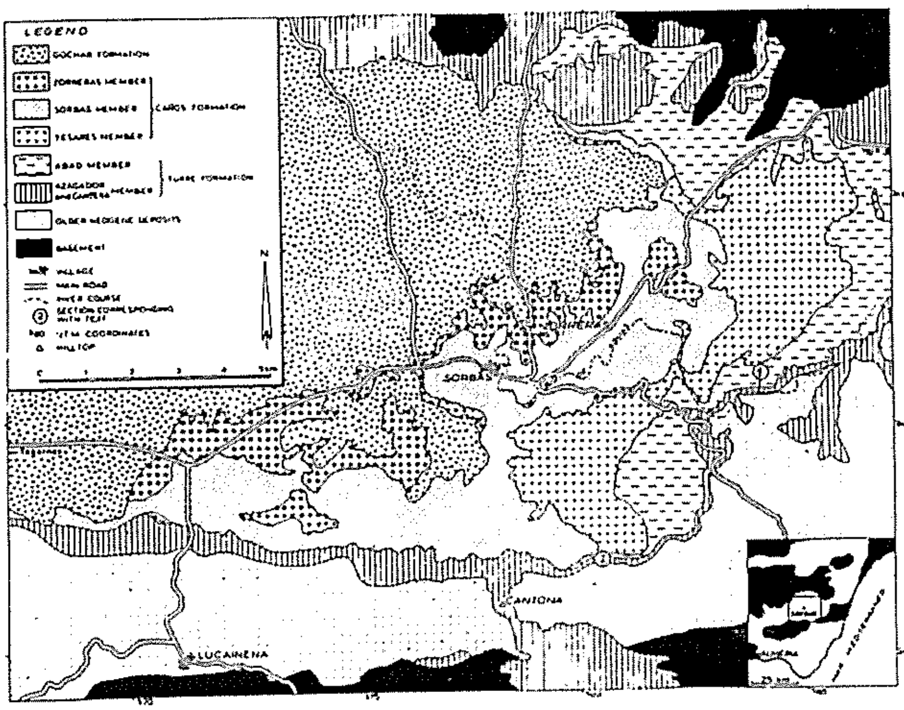

Fig. 1. Geological sketch map of the Sorbas Basin (after Dronkert 1976). 1. Los Perales Section. 2. Hueli Section.

The lower parts of the investigated sections consist of bluish grey hemipelagic marly sediments, slightly sandy at the base, with intercalations of indurated, often silicified levels. The upper part consists predominantly of more or less tobacco coloured very fine-grained sediments, sometimes finely laminated. Diatomaceous intercalations are especially well developed in the uppermost part of the Los Perales Section. Plant remains and thin-shelled pelecypods are often found in the higher parts of both sections. Some predominantly bioclastic calcareous sandstone or conglomerate intercalations present in this interval are thought to represent turbidite ore debris-flow deposits.

In both sections the above described sequence is overlain by gypsum deposits. The uppermost part of the Hueli Section - to a maximum of ten meters - is badly exposed, and was therefore not taken in consideration; the Los Perales Section reaches up to the first gypsum layer.. The last mentioned section is considered to represent deposition in the deepest part of the Sorbas Basin. The Hueli Section, which has a considerably smaller thickness, has a paleogeographic position relatively closer to the basin margin (Fig. 1).

PALEOECOLOGY

1. Benthic foraminifers.

a) Los Perales Section (Figs 2 & 3)

A total number of 94 benthic species was recovered from the Los Perales samples. The majority of these forms are calacreous species; arenaceous species are confined to the lower part of the section. Miliolids are virtually absent. The distribution of a selected number of species is shown in Figure 2A, B, C. Remarkably, only three species - *Cassidulina laevigata, Bulimina elongata* and *Bolivina spathulata* - are present throughout the section. The cumulative percentages of these species (Fig. 2E) shows that the former two species are particularly common in the lower partof the section, while the latter species dominates the benthic assemblages in the upper part. A distinct faunal break is present above sample GG 1040. Figure 2G shows the dominant species in each sample. Fourteen forms characterize the assemblages; only five are dominant in more than one sample. Based on the dominant species, the following benthic zones can be recognized in the Los Perales Section (Fig. 2D):

a: 1) *Bolivina/Cassidulina* zone (GG 1016 - GG 1023).
Other common forms in this interval, besides the nominate taxa, are *Cibicides boueanum, Bolivinoides miocenicus, Uvigerina flintii, Sphaeroidina bulloides, Anomalinacolligera* and small specimens of *Melonis barlaeanus, M. soldanii, Planulina ariminensis* and *Gyroidina sneooldanii.* Species diversity is high (see Fig. 3F). The estimated paleowaterdepth for this part of the section is 200-300 m.

a: 2) transitional zone (GG 1024 - GG 1031).
Six different dominant species chracterize this interval. Common species in this zone are *Cibicides ungerianus, Uvigerina peregrina, Bolivinoides miocenicus* and *Globobulimina* sp.. The paleodepth is thought to be similar to that of the *Bolivina/Cassidulina* zone (200-300 m).

a: 3) *Bulimina elongata/Cibicides pseudoungerianus* zone (GG 1032 - GG 1040).
Common species in this interval, besides the nominate taxa, are *Cibicides boueanum, Valvulineria bradyana, Bulimina striata, Globobulimina* sp. and *Uvigerina peregrina.* Small species of *Melonis* spp., *Gyroidina* spp. and *Planulina* sp. are rare or absent. Species diversity is slightly lower than in the underlying zones. The paleodepth is slightly lower than in the underlying zones. The paleodepth is estimated at around 200 m.

a: 4) *Bolivina spathulata* zone (GG 1041 - GG 1063).
Bolivina spathulata strongly dominates the benthic assemblages in this interval. *Bulimina elongata* and the *Uvigerina cretensis* group occur in minor numbers. Species diversity is low. The paleodepth is estimated at 100 - 150 m. The top of the section contains sparse *Ammonia* sp. and *Elphidium* sp., which, if they are not transported, may indicate slight further shallowing.

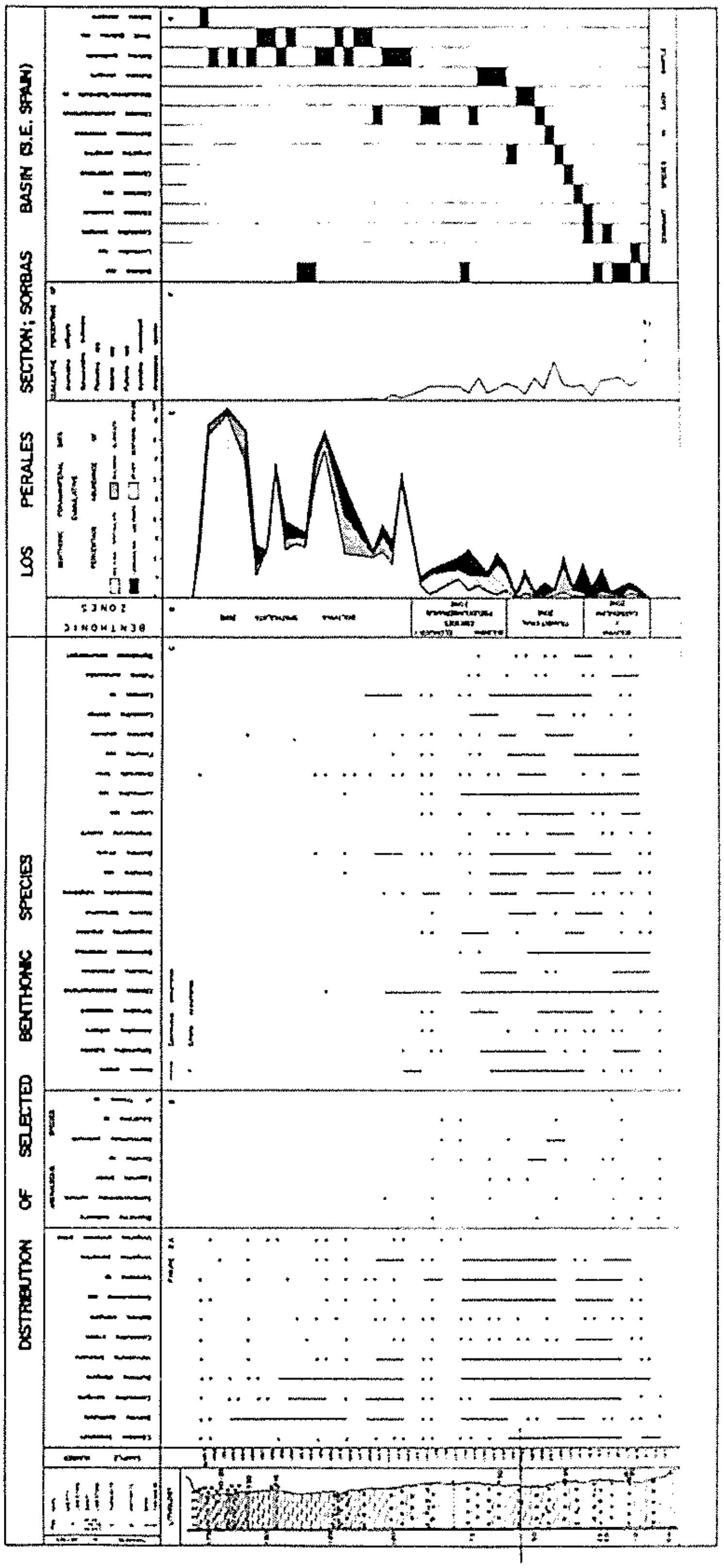

Fig. 2. Distribution of selected benthic species in the Los Perales Section.

b) Hueli Section (Fig. 4)

The samples from the Hueli Section yielded a total of 75 benthic foraminifer species. As in the Los Perales Section, calcareous forms dominate over arenaceous species, which are confined to the lower part of the section. Miliolids are rare to absent. The distribution of a selected number of species is shown in Figure 4A. A faunal break is conspicuous above sample PG 306.

Six dominant species characterize the assemblages (see Fig. 4B). Based on these species four benthic zones have been distinguished (Fig. 4C):

b: 1) lower *Bolivina spathulata* zone (PG 342 - PG 332).
Bolivina spathulata, Melonis barlaeanus, Gyroidina neosoldanii, G. acuta, Pullenia bulloides, Cassidulina laevigata, G. crassa and *Eponides* sp. A are common species in this interval. Species diversity is high (Fig. 4E).

b: 2) *Cibicides pseudoungerianus* zone (PG 330 - PG 320).
The faunal assemblages from this interval are virtually similar to those from the lower *Bolivina spathulata* zone. New in this zone is *Melonis soldanii*, while *Planulina ariminensis* shows a more consistent occurrence.

b: 3) *Bulimina elongata/Cassidulina* spp. zone (PG 318 - PG 306).
Cassidulina laevigata, C. crassa, Bulimina elongata and *Gyroidina neosoldanii* are chracteristic species in this interval. Most of the deeper-water forms disappear or occur in very low numbers above this zone.

Although the dominant taxa in these three zones are different, the benthic foraminifer composition is fairly similar throughout these intervals. The presence of relatively deeper-water benthic species such as Melonis spp., Pullenia spp., Sphaeroidina bulloides, Anomalina colligera and Gyroidina neosoldanii points to paleowaterdepth of 200-300 m for this part of the section.

b: 4) upper *Bolivina spathulata* zone (PG 304 - PG 280).
The nominate taxon makes up almost 90 % of the total benthic assemblages. Associated species are *Bulimina elongata, B. striata, Bolivina* spp., the *Uvigerina cretensis* group and *Cassidulina* spp.. Although most of the deeper-water benthic forms from the underlying zones are absent in this interval, the species present indicate waterdepths of 150-200 m.

In both investigated sections a distinct faunal break is present at the transition from the bluish-grey marls of the lower part, to the tobacco-coloured marls of the upper part of the sections; around GG 1041 in the Los Perales Section, around PG 306 in the Hueli Section (see Figs 2 & 4).

Above these levels, the benhic foraminifer assemblages in both sections are characterized by the following factors:

- arenaceous species are either very rare or absent (Figs 2A & 4A)
- Bolivina spathulata dominates (up to 90 %) the assemblages (Figs 2E, 2G, 4B)
- relatively deeper-water benthic foraminifers such as Anomalina colligera, Melonis soldanii, M. barlaeanus, Pullenia bulloides, P. quinqueloba, Planulina ariminensis, Gyroidina neosoldanii and Sphaeroidina bulloides are either absent or show a sharp decrease in abundance (Figs 2C, 2F, 4A, 4D)
- the Uvigerina cretensis group is subdominant
- red-stained specimens are very common compared with the lower parts of the sections (Fig. 4F)
- species diversity is low (Figs 4A, 4E, 2A, 2B & 2C, 3F).

2. Planktonic foraminifers

As shown on the distribution chart of planktonic foraminifera by Geerlings (1977, see also Troelstra et al. 1979), the *G. scitula* group, the *G. suterae* group and keeled globoratalids, such as the *G. conomiozea/miotumida* (*G. miotumida* = *G. merotumida* of Geerlings 1977) group abruptly disappear above sample GG 1044 of the Los Perales Section. Representants of the *Neogloboquadrina acostaensis* group show a scattered occurrence above this level.

Most persistant species throughout the section are *Orbulina* spp., *G. pseudobesa* and forms of the *G. quinqueloba* group. Monotypic faunas, consisting only of *O. universa* or forms of the *G. quinqueloba* group, are found in some samples from the uppermost part of the section. Forms of the *G. bulloides* group, of the *G. nepenthes* group and *Globigerinoides* spp. are less common, but do still appear in the upper part of the section.

As planktonic foraminifers seem to be especially sensitive to temperature changes, the following analyses were carried out on their assemblages, to determine the effect of this factor operating in this section:

1) a comparison was made between the planktonic foraminifer assemblages occurring in the section and the associations of planktonic foraminiferids in recent marine environments, as described by various authors.

2) three temperature-diagnostic groups were distinguished and their relative frequency percentage plotted (Fig. 3A):

a. a warm-water *G. trilobus/sacculiferus* group. Other characteristic warm-water forms are absent in the Los Perales Section.

b. a temperate/cold water *G. bulloides* and *G. quinqueloba* group.

c. other species with relatively wide or ill-known temperature ranges. This group includes the *N. acostaensis* group, the *G. suterae* group and other forms.

The warm-water *Globigerinoides* group is slightly better represented in the lower part of the section than in the middle and upper parts, above GG 1033. The temperate/cold water *G. quinqueloba* group increases in relative importance from GG 1033 on and is distinctly more common in the upper part of the section. The distribution of the G. bulloides group through the section is fairly similar to that of the *G. quinqueloba* group, except in the uppermost part of the section.

3) the diameter of *Orbulina universa* (Fig. 3B and 3C).
O. universa is present throughout the whole section, mostly in fair abundance. In the Indian Ocean, according to Bé et al. (1973), the diameter of this species decreases towards higher latitudes. Changes in diameter of *O. universa* could thus reflect changes in paleoclimate.

In the Los Perales Section the mean diameter of *O. universa* was measured for 50-75 specimens per sample. Below GG 1043 values fluctuated between 300 and 430 μ, above this level values between 250 and 350 μ were measured. Figures 3B & 3C show the average and maximum diameter values of O. universa, respectively. Figure 3D gives the abundance of *O. universa* as a percentage of the total foraminifer assemblage, indicating that this species is more abundant above sample GG 1043. Similar studies were carried out by Vismara Schilling and Stradner (1977) on samples from the Early Pliocene Trubi Fm at Buonfornello (Sicily). It is interesting to note that their values at the base of the section (360 μ, decreasing to 300 μ in the next sample) are in good agreement with the values obtained from the top of the Los Perales Section (290 μ).

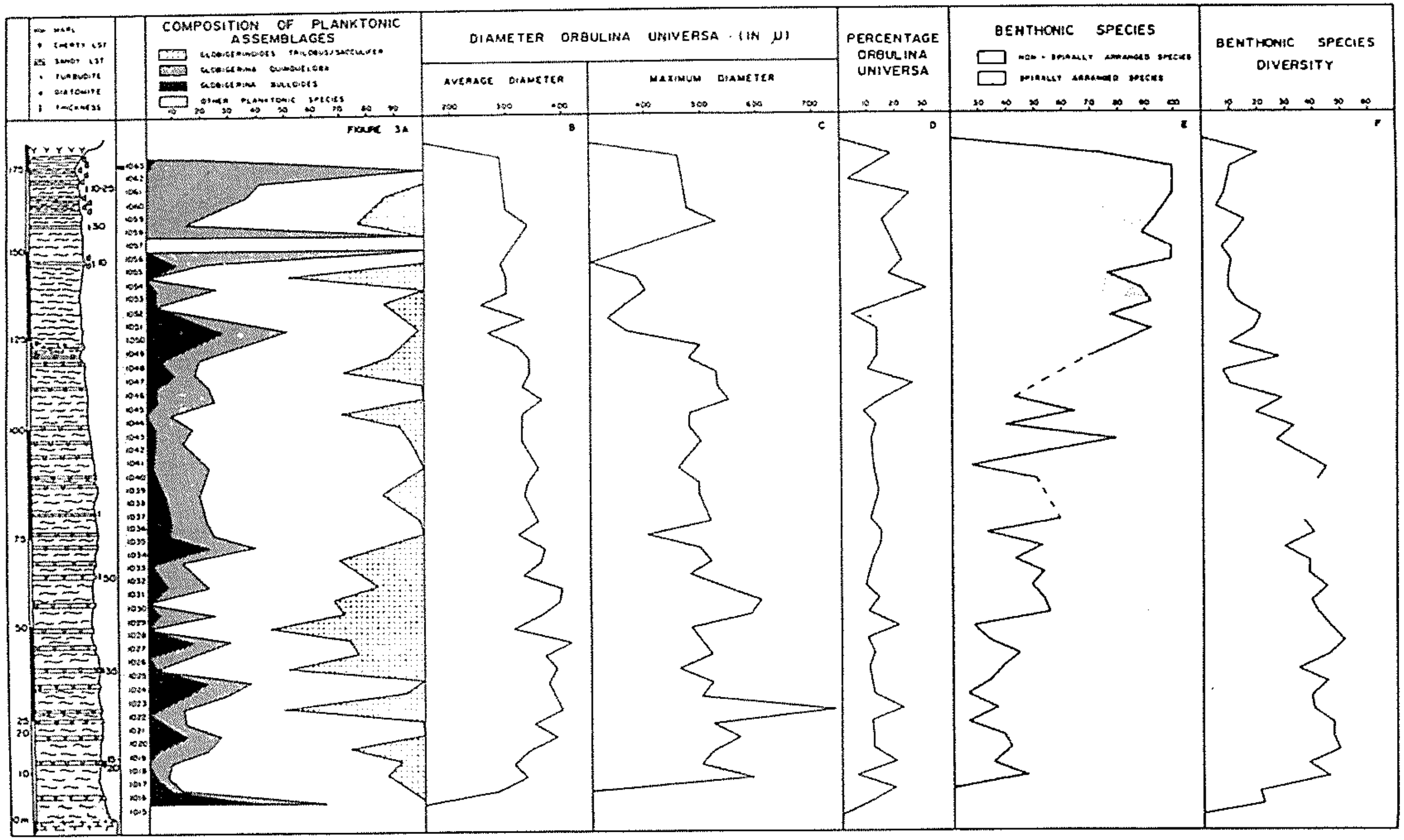

Fig. 3. Composition of planktonic assemblages in the Los Perales Section.

DISCUSSION

Preliminary interpretation of the foraminiferal data from the Los Perales and the Hueli Section seems to indicate that several chemical and physical parameters changed during the deposition of the pre-evaporite sediments in the Sorbas Basin. These will be discussed in the light of our present knowledge of the lithofacies development of the area.

Changes in salinity

The Los Perales and the Hueli Section comprise sequences of sediments, that were deposited before the onset of the main salinity crisis in the Mediterranean. Evaporites are overlying the marl series in the central part of the Sorbas Basin. In the centre of the adjacent basins a similar sequence is found. A restriction of the environment is reflected in the bioclastic or reef deposits at the border of the basin, which contain oligotypical reef faunas, dominated by *Porites*, in their upper parts.

The composition of the benthic foraminifer faunas in the upper part of our sections, is distinctly anomalous. Faunal diversity is low, but the species present indicate a paleodepth exceeding 100 m. At such a depth, rich and diverse benthic foraminifer assemblages would be expected. This indicates extreme bottom conditions, caused by restricted water circulation. These conditions are also indicated by the diatom blooms expresssed in the diatom layers in the upper part of the Los Perales Section. Species of *Bolivina*, *Bulimina* and *Uvigerina* are considered species with a wider tolerance towards extreme environmental conditions (Cita 1973). These forms are the dominant species in the upper part of both sections. taking into account the contemporaneous lithofacies development, it seems reasonable to assume that a gradual increase in salinity played an important role in the restriction of the environment during the later part of the pre-eveporite phase.

The fact that the deterioration of the planktonic fauna seems to set on slightly later, could substantiate the above hypothesis. Surface waters could have remained longer unaffected by the extreme circumstances. Planktonic faunas with extremely low diversities in the uppermost part of this section - monotypical *O. universa* or *G. quinqueloba* assemblages - could indicate a more or less uniform effect on the entire water-column in the latest pre-evaporite phase.

Paleo-waterdepth

An initial transgression can be deduced from the lithofacies development and from the successive benthic and planktonic foraminifer assemblages, which increase rapidly in diversity and change in composition. After this transgression, the overall aspect of the sedimentary sequences in both sections is that of a relatively slowly shallowing upward sequence. A decreasing waterdepth from 200-300 m for the lower part of the sequences to around 150 m for the upper part, is indicated by successive changes in the benthic fauna composition. According to Bé & Tolderlund (1971), the same process might be indicated by:

a) disappearance of keeled globorotalids, *G. scitula* and the *N. acostaensis* group
b) a relative increase in abundance of spinose forms, such as *G. quinqueloba* and *Globigerinoides* spp.

These results are in fairly good agreement with data from Dronkert (1977a), who arrived at a maximum depositional depth of 400 m for the base of the Los Perales Section. His calculations, based on basin geometry, litho- and bio-facies development, sediment

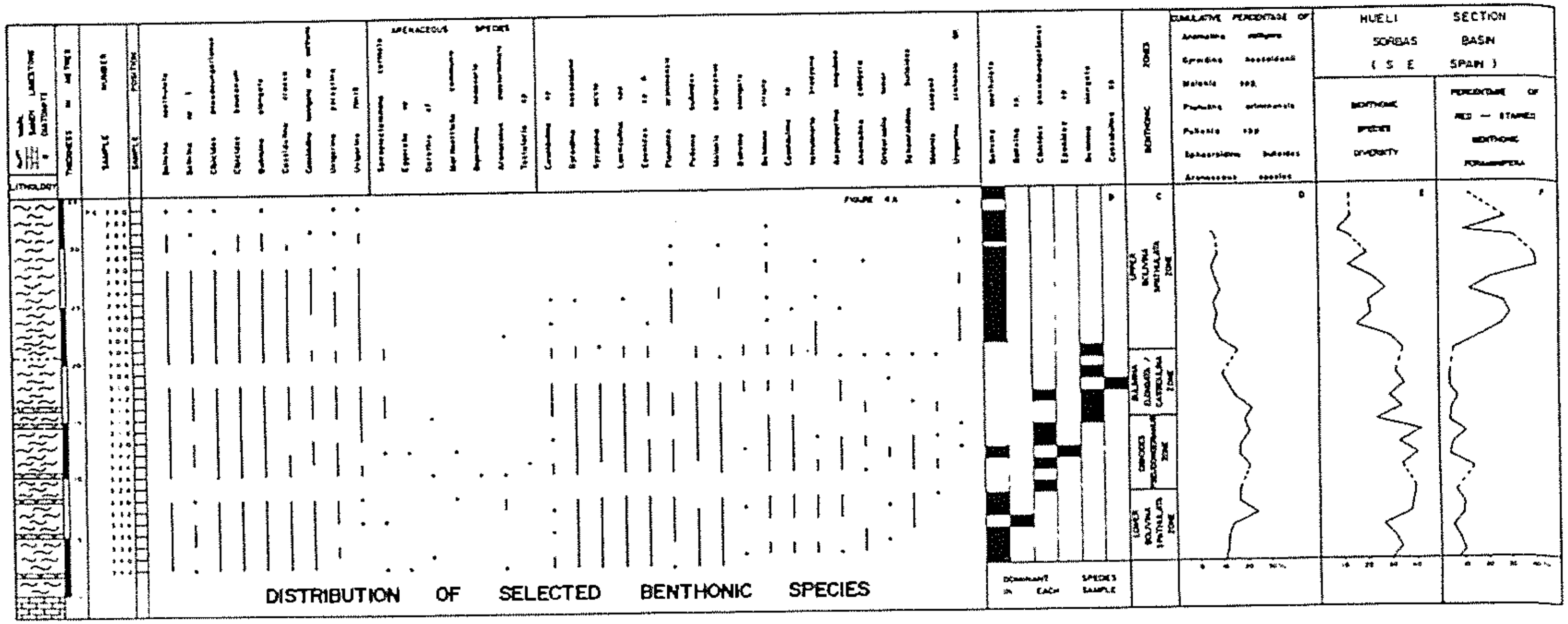

Fig. 4. Distribution of selected benthic species in the Hueli Section.

thickness, tectonics and compaction, resulted in a waterdepth of about 150 m for the top part of the marls in this section. A general shallowing, at least at the basin borders, is also indicated by the outbuilding of reefs in the direction of the centre of the Sorbas Basin towards the end of the Messinian. The same general trend can also be observed in the neighbouring basins (e.g. Vera, Campo de Nijar, Agua Amarga, Almeria and Campo de Dalias Basin).

Considering the position of the Hueli Section, close to the basin border, it is rather surprising that not much difference in paleowaterdepth was found between the sediments of this marginal section and those from the Los Perales Section from the basin centre. This might indicate that the marginal position of the Hueli Section did not become manifest until relatively late in the early Messinian.

Changes in water-temperature

Paleoclimatological data are only available from the Los Perales Section. Faunal data supplied by other authors suggest that the early Messinian pre-evaporite phase in the Mediterranean was a relatively cool period, that started in the Tortonian (e.g. Bizon & Müller 1977). Surface water temperatures of about 16 to 18 °C are suggested by Burckle (1977) from the diatom assemblages in the latest pre-evaporite diatomites of the Sorbas Basin. Planktonic foraminifer data from the Los Perales Section point in the same direction:

a) the absence of warmer-water species (e.g. *G. menardii* s.l.).
b) the presence of cooler-water forms (e.g. *G. conomiozea* group and *G. scitula*).
c) relatively low frequencies of representatives of the *G. trilobus/sacculiferus* group as compared to relatively high frequencies of the *G. bulloides* and *G. quinqueloba* group.

Progressive cooling might be indicated by:

a) the decreasing diameter of *O. universa* towards the top of the section.
b) an increase in relative frequency of spirally arranged benthic forms towards the top of the section.
c) a relative drop in frequency of representatives of the *G. trilobus/sacculiferus* group and a relative increase in frequency of specimens belonging to the *G. bulloides* and *G. quinqueloba* group between the lowermost- and middle part of the section. This latter change takes place around GG 1033, in a sequence where no other lithologic or paleontologic changes seem to occur. Especially *G. quinqueloba* seems to be a good indicator of cooler circumstances. Bradshaw (1959) figures the highest abundance of this species below 18 °C and Bé and Tolderlund (1971) mention even colder (subarctic) conditions (cf. Barash 1971).

Keeled ('menardii-form') globorotalids prefer warmer envirronments (Stainforth et al. 1975) and their absence is generally assumed to indicate water temperature below 17 °C (Bandy 1964). In the Los Perales Section we did not encounter any keeled globorotalids above GG 1044. However, some care must be taken in interpreting a cooler climate for the upper part of the Perales Section. The marked increase in salinity, as well as certain shallowing, may have important effects on the distribution of the Planktonic forms. Similarly, the effect of increased salinity on the diameter of *O. universa* is unknown.

CAUSES FOR THE OBSERVED PHENOMENA AND THEIR IMPLCATIONS FOR THE GEOLOGICHISTORY

The observed restriction of the environment is thought to have been caused by the coming into existence of a sill, separating the area from the open sea. A seemingly contemporaneous restriction of the environment is reported from a great number of circum-Mediterranean secions towards the east of the area under discussion. As continuously open marine sequences, covering the latest Miocene, seem to be present towards the west (e.g; the Guadalquivir Basin), it is most plausible to picture the open sea in that direction. The Sorbas Basin had already become a shallow basin towards the end of the Tortonian. After an initial deepening of some hundreds of meters in the earliest Messinian, a shallowing is visible here, as well as in the surrounding areas. Therefore it seems not unlikely, that the area formed part of a greater Betico-Riffian barrier system, separating during the Messinian the deeper Mediterranean basins (see Hsü et al. 1977) to the east, from the Atlantic in the west. If the observed trend deduced from our data, may be extrapolated, sedimentary infilling of already shallow basins, together with obstructing reef-growth, may have contributed to a further shoaling of the area during the Messinian. This may have been a factor in generating a sufficiently shallow barrier between the Atlantic and the Mediterranean, to cause the 'Messinian Salinity Crisis'. The fact that planktonic faunas from the Sorbas Basin seem to indicate a relatively cool climate during the earliest part of the Messinian, and the possibly superimposed cooling trend, could lead to the conclusion, that eustatic sealevel lowering caused by growing ice-caps, may have contributed to the shoaling effect.

Evidence for tectonic movements in the SE part of the Province of Almeria, can be found in sediments of latest Tortonian age. A similar feature is reflected in the deposits from the time-interval around the Mio/Pliocene boundary (e.g. Iaccarino et al. 1975, Montenat et al. 1976). These tectonic processes may also have been a cause for shallowing.

However, more work has to be done to unravel the problem of the relative importance of the above-mentioned processes and their exact time relationship.

Judging from the litho- and biofacies development of the area under discussion, as well as its paleogeographic position, it seems that the main axis of the barrier system was situated outside it, and probably more towards the west. Although more exact information is still needed on the facies relationships, both within and between the different basins, this conclusion seems to be substantiated by data from other basins in the Betic area.

CONCLUSIONS

An initial rapid transgression took place during the latest part of the Tortonian, and the earliest part of the Messinian. At a relatively early moment during the Messinian pre-evaporite phase, restriction of the environment took place in a rather shallow (max. a few hundred meters waterdepth) Sorbas Basin, during a relatively cool period. A fluctuating, but progressively increasing salinity is thought to have played at least an important role in the environmental restriction. Due to this restriction, other environmental factors have to be considered very carefully. It seems probably however, that this restriction was preceeded and accompanied by a certain shallowing of the area, which is also apparent from the lithofacies development. A slight cooling trend might have been an accompanying factor too.

The main cause of the observed environmental changes seems to be the coming into existence of a sill, separating the area from the open sea. It was probably situated to the west of the area, obstructing its free connection with the Atlantic. Several causes for the genesis of this barrier have been considered on basis of the data presented. Amongst these are: sedimentary infill of shallow basins, obstructing reef growth, eustatic sealevel lowering due to the growing of the Antarctic ice sheet and tectonic processes in the Betic area during the latest Miocene.

More accurate investigation of the litho- and biofacies development of the different basins in the Betic realm is however needed to conclude on the relative importance of these processes and their exact timing.

4.4.1. PRELIMINARY RESULTS OF A BIOSTRATIGRAPHIC STUDY OF THE MIO/PLIOCENE BOUNDARY IN THE PROVINCE OF ALMERIA (S.E. SPAIN)[1]

L.P.A. Geerlings & H.M. van de Poel

Based on a study of literature and our own research, we selected the most 'unambiguous' biostratigraphic (planktonic foraminifer and nannoflora) markers for usage in the Mediterranean and adjoining Atlantic area (indicated with an * in Fig. 1). The others (b, d, e) atre considered moderately reliable marker-levels. Their 'ideal' succession and their most probable relation to the paleomagnetic time scale is figured on the left, their general succession and relation to the Mio/Pliocene boundary in the Mediterranean on the right.

Application of these datum levels on sections from the central parts of some basins in S.E. Spain, has led to the following preliminary conclusions:

a) The 'early Messinian interval' (a*-c*) is commonly well developed in all the studied basins.

b) The 'late Messinian-early Pliocene interval' (c*-f*) is generally poorly represented biostratigraphically (Sorbas, Campo de Nijar, Agua Amarga, some sections in the Almeria Basin), because of hiatuses, oligotypical development or absence of fauna, due to abnormal marine conditions.

c) In the Vera Basin, interval c*-f* seems best developed at first sight, since most marker levels (their spacing is represented roughly in meters of sediment in the centre column) are present in the same succession as 'ideally' expected. Yet an anomalous situation is found: marker levels b, c* and d in the nearby Atlantic, and b and c* in the Mediterranean occur some hundreds of meters below the Mio/Pliocene boundary (representing several hundred thousand years), whereas in the Vera Basin these markers occur practically at the Mio/Pliocene boundary (as it is assumed by most authors, see arrow), in a few meters of sediment.

This situation points strongly to either extremely condensed sedimentation (with reworking of older markers b and c* in an abnormal marine environment), or to the presence of an hiatus.

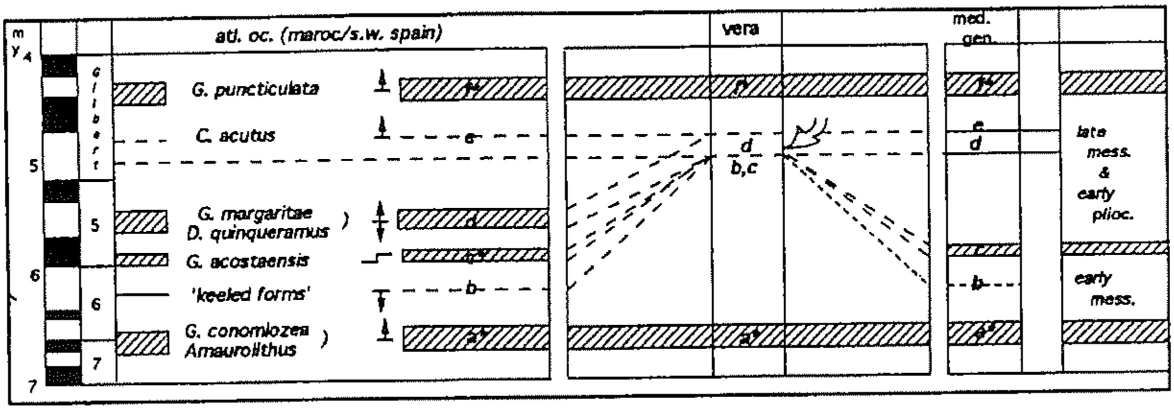

Fig. 1. Position of Mio/Pliocene 'marker levels' in the open Atlantic (left), the Mediterranean (right) and the Cuevas del Almanzora Section of the Vera Basin (SE Spain, centre).

[1]Text and figure from *J.M. Rouchy & F. Orszag-Sperber (eds.): Résumés des communications, Vème Séminaire sur le Messinien (Paphos, Chypre, 1979). UNESCO/IUGS (IGCP Proj. 117).*

4.4.2. CHARA SP. IN MIO-PLIOCENE MARLS AT CUEVAS DE ALMANZORA, VERA BASIN, S.E. SPAIN.[1]

L.P.A. Geerlings, H. Dronkert, H.M. van de Poel & J.E. van Hinte

Summary

This note reports the finding of non-marine plants in a varicolored marl section at Cuevas de Almanzora in the Vera Basin. The section previously has been interpreted as an uninterrupted marine Upper Miocene-Lower Pliocene succession. The preservation of undamaged, fragile Chara sp. branchlets with oogonia excludes reworking of those fresh water plants and, together with the presence of laminae with abundant delicate valves of Cyprideis sp. of all mold sizes, evidences that the marls have a non-marine origin.

The section no longer can be considered to be a continuously marine record of the Mio-Pliocene transition. An inferred erosional period most likely corresponds with the time of lowered sealevel and evaporite deposition elsewhere in the Mediterranean Basin.

INTRODUCTION

The 'classical' Cuevas de Almanzora section (Fig. 1) was briefly examined in June 1979 to study its feasability for geomagnetic analysis. The succession described by Montenat et al. (1976) could be recognized showing three units (Fig. 1, Plate Ia):

unit A, a Miocene sequence of buff marine marls alternating with thin turbidites, overlain by

unit B, varicolored micaceous laminated marls with some sandy stringers, in turn overlain by

unit C, Pliocene buff marls with thin sand beds

Unit B encompasses strata which have been interpreted as a continuous marine record of the Mio-Pliocene transition (Montenat et al. 1976).

DESCRIPTION

Open marine planktonic and benthic foraminifers can be seen with a hand lense in all these marls, whereas those of Unit B in addition display non-marine fossils. Smooth and (fewer) pustulose) valves of different mold sizes of the non-marine ostracode *Cyprideis* sp. are common throughout the unit and can occur in great abundance (Plate 1b). We also found scattered thin-shelled casts of gastropod shells.

Most significant, though, many bedding planes of unit B have common brown plant remains among which up to several dm. long fragments of grass and/or reed and

[1]Text and figures earlier published in *Proceedings of the Koninklijke Nederlandse Akademie van Wetenschappen, Series B, Volume 83 (1), (1980), pp. 29-37.*

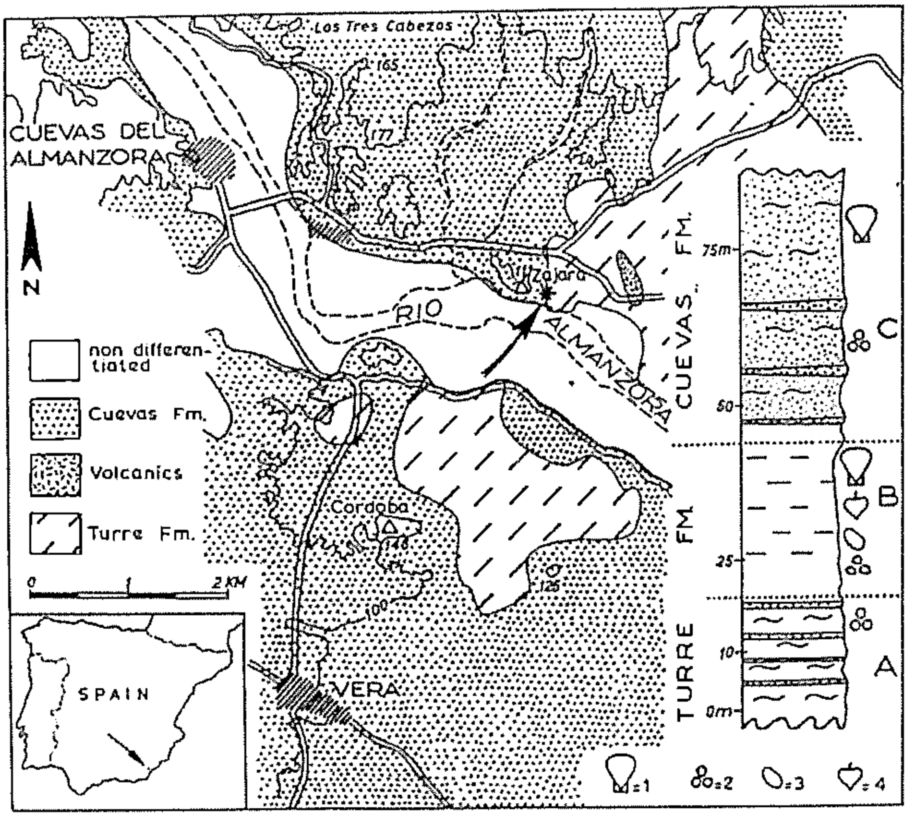

Fig. 1. Geological setting of the Cuevas del Almanzora Section. (Map after Völk 1967). 1. Pelecypods. 2. Foraminifers. 3. Ostracodes. 4. Plants. * = section location.

Characeae are the most conspicuous. On freshly cut surfaces the *Characeae* are well preserved and show the articulate nature of their main branch and branchlets (Plate Ic). Several speci-mens were found with oogonia preserved in the axillae of the branchlets (Plate Ic, d).

Only few oogonia have details of the spiral structure preserved but, once we had recognized them in the axillae, the use of a hand lense sufficed to find free specimens throughout the marls. Particularly well-preserved *Chara* were recovered from near the top of the unit in a freshly dug water reservoir behind the traditional riverbank exposure (Plate Ia). Montenat et al. (1976) also mention the occurrence of 'worn and reworked' oogonia of *Chara* in the middle part of unit B.

Laboratory study of our sample collection confirmed the observation by Montenat et al. (1976) that a well preserved planktic foraminiferal fauna is present throughout the section, and assemblages of Late Miocene age (in units A and B, with *Globorotalia mediterranea* Catalano and Sprovieri) and Early Pliocene age (in unit C, with *Globorotalia margaritae* Bolli and Bermudez) could be distinguished indeed. Other than Montenat et al. (1976), however, we found in unit B an irregular alternation of sinsitral and dextral populations of *Neogloboquadrina acostaensis* (Blow) which is anomalous. Dextral populations occur as low as level 74150 of Montenat et al. (1976). At that level they recorded a sharp break in clay composition signifying the onset of rapid hinterland erosion (Montenat et al. 1976, figure 6, p. 634). Carbonnel (in Montenat et al. 1976, p. 629-630) reports marine ostracodes, *Cyprideis* sp. and lacustrine *Tyrrhenocythere* and Cyprididae from unit B.

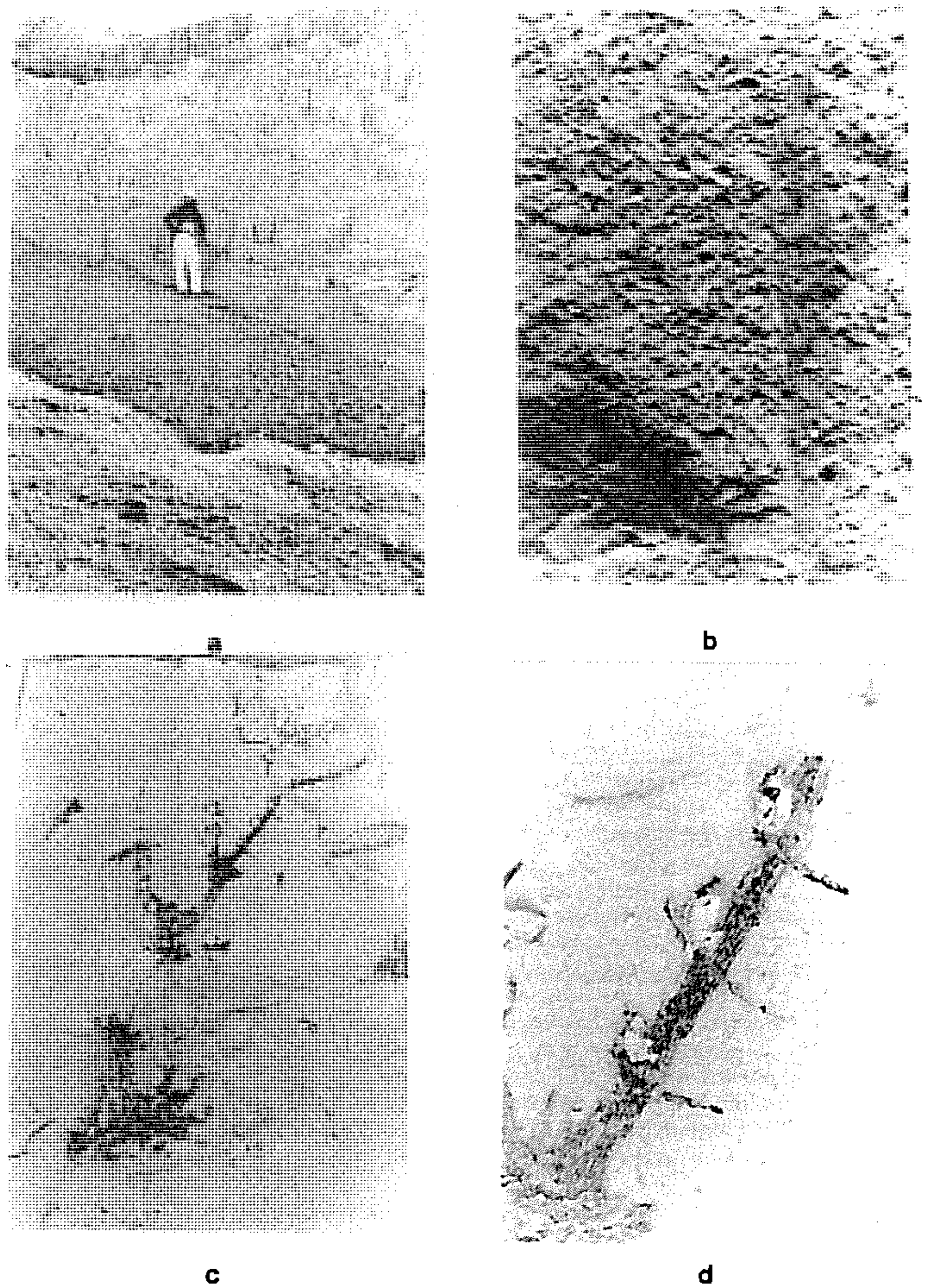

PLATE 1

a. Photograph showing freshly exposed section in waterreservoir behind riverbank outcrop.
b. Macrophotograph showing abundant ostracodes (*Cyprideis* sp.) on bedding plane. Note different mold sizes and reed fragment. x 2.
c. Macrophotograph showing bedding plane with *Chara* specimen. Note the articulate stem and branchlets with oogonia preserved in the axillae. x 1.5.
d. Detail of macrophotograph c. Oogonia are about 1.45 mm long and 1.10 mm wide and possess 8 to 12 spiral whorls. x 8.5.

DISCUSSION

The occurrence of the well preserved non-marine fossils described above is evidence that the marine elements were redeposited in a relatively shallow 'fresh water' lake. In fact, the anomalous nature of the marine assemblage (Dronkert et al. 1979, Geerlings & Van de Poel 1979), supports this idea, as do the clay mineralogy data of Montenat et al. (1976) and the calcareous nannofossil data of Backman (1978). Maintaining that the planktonic faunal association is a natural one would imply that the reference scheme, as established elsewhere in the Mediterranean and the Atlantic, is at err; the *Cyprideis* sp. valves being displaced into a normal marine environment and the repeated co-occurrence of different mold sizes being an accidental insensitivity for sorting. However, now with the record of undamaged *Chara* sp., there is no further doubt about the non-marine nature of unit B and despite its good state of preservation the plankton can safely be considered to be reworked and mixed.

At the earliest, deposition of this interval took place late in the Miocene, and the lake most likely was part of the late Messinian 'Lago Mare' (cf. discussion by Benson 1976, 1978a and Cita et al. 1978). The deposits could well be a more basinal equivalent of at least part of the coastal-terrestrial sequence described by Roep and Van Harten (1979). Consequently there would be a significant unconformity (non-depositional or erosional hiatus) between units A and B: uppermost Messinian 'Lago Mare' deposits directly overlying lowermost Messinian beds (co-occurrence of *Globorotalia mediterranea* and sinistral *Neogloboquadrina acostaensis* indicates *Globorotalia mediterranea* Subzone of D'Onofrio et al. (1975), and lower part of *Globorotalia conomiozea* Zone of Van Hinte et al. (in press). During our reconnaissance study we could not discern a hardground, condensation horizon or other indications for non-deposition and therefore we think it likely that the hiatus is erosional. It will be interesting to test the present time-stratigraphic conclusions by comparing the results of a paleomagnetic analysis with the polarity reversal scale.

Considering the topographic position of the section and the fact that it lies between older and younger 'deeper' marine deposits, the most plausible explanation for erosion seems to be a drastically lowered sealevel. Considering its geographic position and age, the erosional period most likely corresponds with the time of lowered sealevel and evaporite deposition elsewhere in the Mediterranean Basin. The surface of the subsequent Lago Mare probably still was well below the original sealevel, for the presence of *Chara* sp. and *Cyprideis* sp. indicates shallow water, and older marine Messinian must have been exposed as it was washed into the lake.

CONCLUSIONS

The Cuevas del Almanzora section contains a record of early Messinian marine deposition, probably followed by erosion, and subsequent late Messinian deposition in a non-marine environment, followed by a return to marine depositional conditions in the Pliocene. This sequence of events confirms rather than negates the hypothesis that the Mediterranean did dessicate in latest Miocene time.

Through combined detailed sedimentologic, paleomagnetic and micropaleontologic analysis of this 'unique' section it will be possible to attain an understanding of the duration and magnitude of the processes and events associated with the Mediterranean 'salinity crisis'.

4. 5. A REMARKABLE LIMESTONE BRECCIA AND OTHER FEATURES OF THE MIO-PLIOCENE TRANSITION IN THE AGUA AMARGA BASIN (S.E. SPAIN)[1]

H. M. van de Poel, Th. B. Roep & N. Pepping

SUMMARY

Sedimentary facies and stratigraphic relationships of the Agua Amarga Basin have been investigated in order to obtain more data on its young-Neogene geologic history. It appears that a transgression led to an open-marine environment, with outer-shelf to uppermost bathyal waterdepths, during the earliest Messinian. This was followed by a period of deposition of sediments indicating restricted marine and hypersaline environmental conditions. A coarse and massive brccia lies on top of these deposits in central parts of the basin. This breccia most likely resulted from a collapse of the virtually azoic limestone material of which it is composed, due to the dissolution of salts that were originally deposited with it.

The sediments of the first depositional cycle are truncated by an important erosion surface that locally has the form of a wide valley and is further characterized by karst and soil development. In the Pliocene there followed a new transgression and open-marine sediments were again deposited in a smaller basin having a shallower maximum waterdepth than that of the early Messinian.

The regressive and erosive conditions that prevailed in between the two transgressive periods are considered to have been effected by the combined processes of local tectonic uplift and widespread sealevel lowering.

RESUME

Les faciès sédimentaires ainsi que les relations stratigraphiques du Bassin d'Agua Amarga ont été étudiés dans le but d'obtenir plus de connaissance sur son histoire durant le Néogène Supérieur. Au début du Messinien, il semble qu'une transgression créa un environnement de mer ouverte à paléobathymétrie carcatéristique des domains néritique externe à bathyal supérieur. Cette transgression fut suivi par une période de sédimentation indiquant un environnement marin confiné et hypersalin. Une brèche grossière et massive recouvre ces sédiments au centre du bassin. Très probablement, cette brèche est le résultat de l'effondrement des calcaires presque azoïques qui la composent, après la dissolution des sels déposés simultanément avec ces carbonates.

Les sédiments du premier cycle sédimentaire sont coupées par une surface d'érosion importante. Localement, cette surface prend l'aspect d'une large vallée et se caractérise également par la présence de karst et des paléosols. Durant le Pliocène, une nouvelle transgression déposa à nouveau des sédiments de mer ouverte, dans un bassin plus petit et à bathymétrie moins profonde par rapport à celle du Messinien inférieur.

Les conditions régressives et érosives qui ont prévalu sur la période intermédiaire entre les deux périodes transgressives, sont considérées comme le résultat combiné de deux processus: la surélévation tectonique locale et un abaissement important, au point de vue régional, du niveau de la mer.

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[1] Text and figures earlier published in *Géologie méditerranéenne 11, 3, (1984), pp. 265-276.*

INTRODUCTION

The subject of this paper is the geologic history of the Agua Amarga Basin, the southeasternmost Neogene basin of Spain, focussing on the time-interval around the Mio-Pliocene boundary. For this purpose, the facies of its Messinian to Pliocene sediments and their stratigraphic relationships have been investigated.

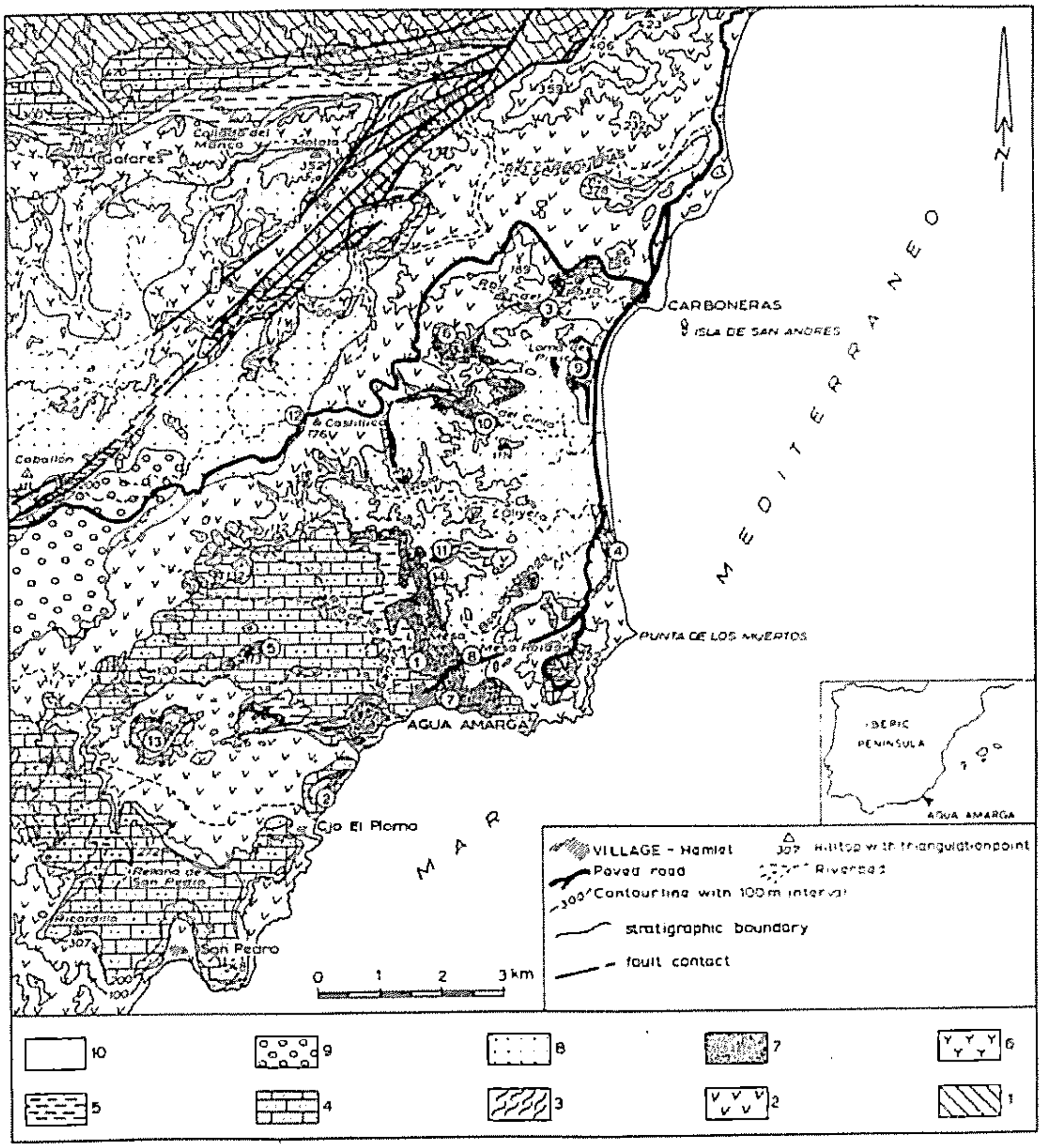

Fig. 1. Simplified geologic map of the Agua Amarga Basin and part of the adjoining Almeria-Nijar Basin with position of localities referred to in the text (encircled numbers).
Legend: 1. Pre-Neogene alpine basement; 2. Volcanic basement; 3. Late Burdigalian-Tortonian sediments; 4. Marginal limestone facies of the Gafares Group (mainly Azagador and Cantera Members of the Turre Formation); 5. Basinal facies of the Turre Fm (Abad Member and reef-derived gravity flows of the Cantera Mb); 6. Caños Fm of the Almeria-Nijar Basin (Messinian gypsum with associated Manco and Feos deposits); 7. Caños Fm of the central part of the Agua Amarga Basin (Agua Amarga Breccia with some associated rock of the Manco and Feos Members); 8. Marine Pliocene (Molata Formation); 9. Campo de Nijar Formation; 10. (Sub)-recent beach gravels and sands.

Provisional descriptions of the sedimentary rocks of the area have previously been given by authors who concentrated on its older, volcanic rocks or its 'neotectonics' (e.g. Coëllo & Castañon 1965, Fúster et al. 1967, Sánchez Cela 1968, c.q. Bousquet et al. 1975a, Bousquet& Philip 1976), or by others dealing with the Neogene sediments on a broader scale (Addicott et al. 1978, Dronkert et al. 1979, Estebán 1979). In recent years, attention has been especially directed towards the characteristic Messinian reef sequences that are developed along the basin margins. This has resulted in a number of, generally increasingly detailed, descriptions (Roux & Montenat 1977, Dabrio & Martin 1978, Armstrong et al. 1980, Estebán & Giner 1980).

We have concentrated on the facies development in the more central part of the basin. In this, emphasis has been laid upon the investigation of the nature, stratigraphic position, and possible origin of a peculiar limestone breccia, having a relatively wide horizontal extension and locally reaching a thickness of up to 40 meters. It is a rather unique phenomenon, as other basins in southeastern Spain with a more complete young Neogene sequence lack such a breccia. Rather, such basins contain Messinian evaporites in a comparable stratigraphic setting. Furthermore we paid some special attention to the relative importance of, and possible cause for, an erosional surface that occurs in between the breccia and subsequent transgressive marine Pliocene deposits, as well as to the general bathymetric evolution of the basin in the period covering the Mio-Pliocene transition interval.

SETTING OF THE AGUA AMARGA BASIN

The Agua Amarga Basin lies nestled in the volcanic rocks that are the northeastern continuation of the Sierra de Gata mountain range. The northwestern margin of the basin is defined by a relatively low range of hills that forms the watershed between the Agua Amarga Basin, in which the drainage pattern is directly eastward, and the northeastern part of the adjoining Almería-Nijar Basin, which, by means of the Rio Carboneras and its tributaries, drains in a more northeastern direction. This landform will be called here the Castillíco Ridge, after a conspicuous hilltop in its central part, just south of the Carboneras-Almería road (see map). To the east, the Agua Amarga Basin is now almost completely open towards the Mediterranean. The Mesa Roldán and, on a much smaller scale, the tiny Isla de San Andres in the bay of Carboneras, form the only interruptions.

The sedimentary rocks that will be discussed more closely below, were deposited upon a volcanic basement. The area was dominated by volcanic activity during major part of the Miocene (c.f. Fúster et al. 1965, Bellon et al. 1976).

DESCRIPTION OF THE SEDIMENTARY FACIES AND THEIR STRATIGRAPHIC RELATIONSHIPS

No formal lithostratigraphic scheme has previously been given for the sedimentary rocks of the Agua Amarga Basin. We will follow, in principle, the pre-existing nomenclature for comparable sediments of the northeastern part of the adjoining Almeria-Nijar Basin (in particular as developed by Völk & Rondeel 1964, Dabrio & Martin 1978, etc.). A number of additional rock units will be defined and described below.

A. Gafares group

This unit comprises a wide variety of facies, that are however all considered to belong to one basic sedimentary cycle. The name is derived from a small village in the northeastern part of the Almeria-Nijar Basin, in the vicinity of which it is well developed (ca. 37° 01' N, 2° 00' W; see map).

In the Agua Amarga Basin, the Gafares sequences are transgressive upon the volcanic basement. The most complete and instructive sections for the development of this unit are here found in the area directly to the north of Agua Amarga (see map and Fig. 2). The following subdivision has been made:

1. *Turre Formation*

The lower part of the Gafares sequence is built by calcareous clastic rocks, reef limestones and fine-grained marly sediments, all rich in marine fossils; These rocks are largely comparable to those of the Turre Formation as it has been described from the nearby Vera Basin by Völk & Rondeel (1964), Rondeel (1965) and Völk (1967). The rocks from the area under discussion have been asigned to the following three classical Members of this Formation (essentially from old to young, lateral relationships are however present):

a. Azagador Member

Orange to yellow bioclastic calcarenites predominate. Conglomerates with well-rounded volcanic pebbles and boulders derived from the underlying basement, bedded in a calcareous matrix containing thick-walled *Ostrea* sp., are developed at the base.
The major development of Azagador rocks, up to several tens of meters thick, is found in the southern-central part of the basin (e.g. localities 1, 2 and 5).

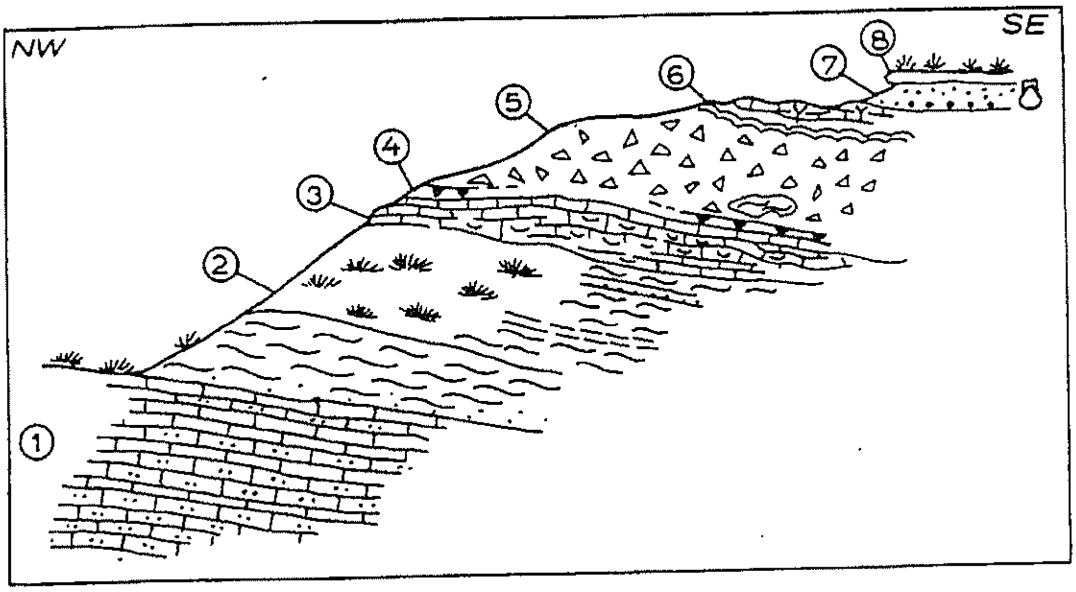

Fig.2. Field-sketch of locality 1 (La Mesa Section).
1-8: Units of the Gafares Group: 1 = Calcarenites of the Azagador Member; 2 = Marls, with diatomite and calcarenite intercalations, of the Abad Member; 3 = Reef-derived debris-flows of the Cantera Member; 4 = Limestone beds, containing some gypsum ghosts, of the Manco Member; 5 = Agua Amarga Breccia; 6 = Reddish limestone, containing knobbly laminated levels of the Feos Member; 7 = Marine Pliocene conglomerates and sands of the Molata Formation; 8 = Caliche crust.

b. Abad Member

Marls that are rich in microfossils typically form the bulk of this unit. Calcareous clastic and diatomitic intercalations are present. The bioclastic beds generally show intense burrowing and grading. They occur more frequently in the upper part of the Abad Member, where they tend to be coarser grained and often contain shells lying with their concave sides upwards.

The bluish-grey marls that dominate the lower part of the Abad member contain a relatively diversified foraminiferal microfauna. Forms of the G. *conomiozea-mediterranea* group are often present within the planktonic assemblages, that are comparable to those described from the lowermost Messinian of Italy by i.a. d'Onofrio et al. (1975). The upper part of the unit, in which the marly sediments often exhibit a somewhat darker color and lamination, contains an oligotypical microfauna. During deposition of these rocks a change in coiling ration of the planktonic foraminifer *N. acostaensis* took place from predominantly sinistral to dextral. We use this event to make an informal subdivision in early, and younger Messinian age for the rocks of the area under discussion (Montenat et al. 1976; cf. Geerlings & Van de Poel 1979).

Abad rocks are mainly developed in the central part of the basin, where they rest with a transitional contact upon sediments of the Azagador Member and reach a maximum thickness of about 30 meters (e.g. localities 1 & 3). Towards the south and southeast, the Abad marls start to interfinger with, and eventually laterally wedge out into, directly overlying clastic Azagador and Cantera sediments (see below), deposited at the foot of the highest, Messinian reef-bearing, paleoslopes of the basin (e.g. localities 2, 5, 13, 4 and 7).

c. Cantera Member

Calcarenitic and calciruditic layers, consisting of reef-derived debris, are the main constituents. They are often crudely graded and sometimes somewhat slumped and contorted. Apart from many shells and remains of other marine organisms, these beds often contain blocks of *Porites* limestone that can range in size from a few decimeters to several meters in maximum diameter. Coral-reef limestone with Porites in life position locally occurs above the highest parts of the irregular volcanic subsurface of the basin (in Mesa Roldán and Rellána de San Pedro, see map; cf. Dabrio & Martin 1978, Esteban & Giner 1980).

Cantera rocks only occur along the southern, and locally, eastern margin of the basin. In sections nearer to these margins, they rest directly upon sediments of the Azagador Member (e.g. localities 2, 4 and 13), and can attain a thickness of several tens of meters. Nearer the basin axis, a few to about ten meters of Cantera debris flows covers marls of the Abad Member (e.g. localities 1, 5 and 9). Northwestwards, the reef-derived debris starts to interfinger with, and eventually laterally wedges out into, marly sediments of the upper part of the Abad Member of the central part of the basin. This lateral facies transition is well exposed in the area to the northeast of the Rambla Viruegas (around locality 1).

It can be remarked that major part of the 'Reef Complex', that has been described by Dabrio & Martin (1978) and Esteban & Giner (1980), is here assigned to the Cantera Member.

2. *Caños Formation*

The younger Gafares sediments have been assigned to the Caños Formation, originally described from the Sorbas Basin by Ruegg (1964, in Völk 1967, Dronkert & Pagnier 1977, etc.). The rocks from the area under discussion are comparable to those of the type area in age, stratigraphic position and also, to a certain extent, in facies. They are for instance similarly characterized by a general scarcity in marine fossils. Since, on the other hand, the facies development in the Agua Amarga Basin displays certain particularities, the following three 'new' Members are recognized here:

a. Manco Member

This rock sequence mainly consists of an alternation of limestones and marly sediments. It is in many aspects comparable to the typical 'Calcare di Base' from the Messinian of Italy. Our name for it is derived from the 'Collado del Manco4, situated about 8 kilometers to the northwest of Carboneras (see map, ca. 37° 01' N, 1° 58' W).

Grey to yellowish, massive, breccious or 'wavy laminated' limestones, in layers of a few centimeters to several decimeters thickness, are the most characteristic of these unit. They often contain irregularly formed small cavities that sometimes take on the shape of gypsum crystals and appear to be mostly azoic, except for the wavy laminations that are reminiscent of the presence of algae. The limestone beds alternate with silty marls which are often laminated.

Only successions of less than one, to a few meters thickness are locally found in the Agua Amarga Basin. These were deposited in its deeper parts, where they now rest conformably upon marly sediments of the top of the Abad Member or upon laterally equivalent Cantera reef-debris deposits (e.g. localities 1, 3, 6 and 7).

Remark: The top of our Gafares sequences of the topographically highest, Messinian reef-bearing parts of the Agua Amarga Basin is locally built by a thin series of limestone in which oolitic beds dominate. This has been more thoroughly described by Esteban & Giner (1980) under 'Terminal Carbonate Complex', of which it builts the main part. It may constitute a lateral equivalent to the Manco sequences of the deeper parts of the basin. A possible assignment to the 'Sorbas Member' of the Caños Formation has previously been suggested by Dabrio & Martin (1978).

b. Agua Amarga Breccia Member

This unit has been named after the small village in which surroundings it is well developed and exposed (see map, ca. 36° 57' N, 1° 56' W). Locality 7, in the cliff at the northeastern end of the beach of Agua Amarga (Fig. 3), has been chosen as type section. The Breccia here reaches its maximum thickness of about 40 meters.

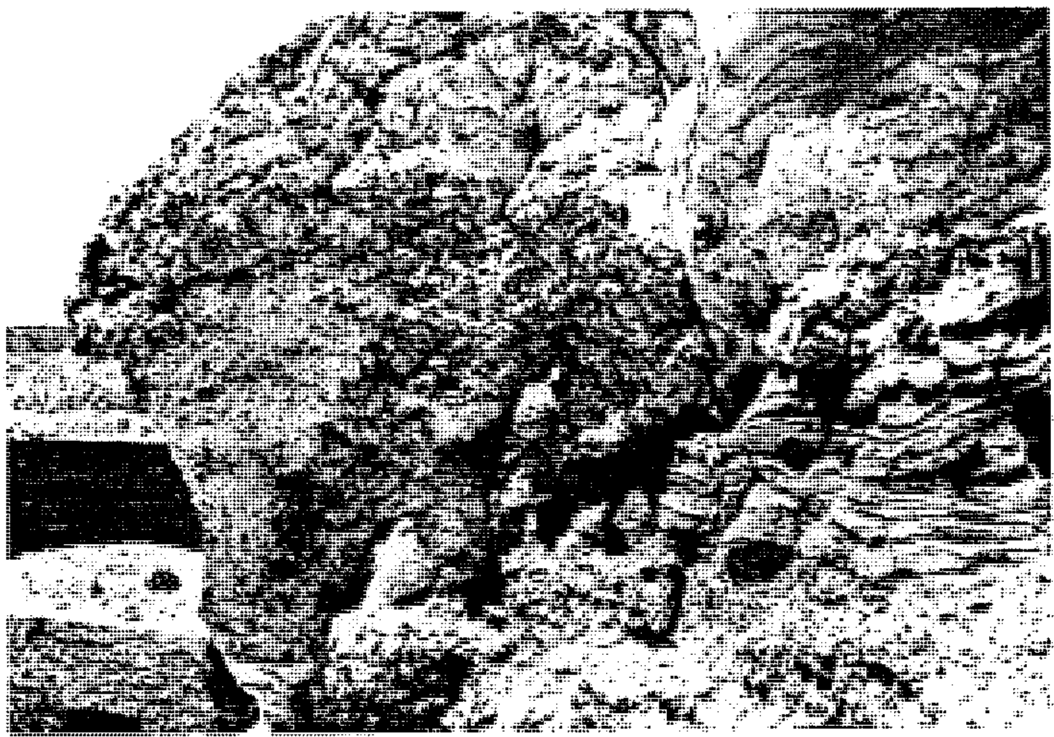

Fig. 3. Aspect of the Agua Amarga Breccia as photographed in locality 7, bay of Agua Amarga. A large slab of thin-bedded limestone appears on the right hand, above the hammer-shaft.

The Agua Amarga Breccia Member wholly consists of one massive braccia whose main constituents are angular fragments and up to 1 meter big blocks of limestones that are similar to those described earlier from the Manco Member. In addition, limestone fragments with cavities in the form of bivalve and gastropod shells have been sometimes found. In a number of sections, contorted slabs of thin-bedded limestone, reaching a maximum diameter of several meters, are present in the lower part of the unit (e.g. localities 1, 5 and 7, see Fig. 3). The massive, chaotic aspect of this Member is due to the lack of any trace of bedding and its ill sorting. The coarser fragments occur in a matrix of finer sand to silt size, consisting of angular to subrounded material of the same nature.
The Agua Amarga Breccia only occurs in central parts of the basin, where it lies on top of the Abad member or laterally equivalent cantera deposits (localities 2, 5, 7, 10 and 11). Often a thin series of Manco limestone is found at the base of the Breccia (localities 1 and 3).

Some occurrences of 'Terminal Carbonate Complex' rock, that Esteban & Giner described from 'lower depositional sites' around Agua Amarga, and in which they observed 'intense brecciation', are assigned to the unit under discussion.

c. Feos Member

The name is derived from a hamlet in the northeastern Campo de Nijar, the Cortijada de los Feos, where this unit reaches a maximum development and is well exposed (ca. 37° 01' N, 2° O2' W). Rocks assigned to this unit occur locally in central parts of the Agua Amarga Basin, where they reach a relatively small maximum thickness of a few meters. The different lithologies here found are also present within the varied facies of the type-area and occur in a comparable stratigraphic setting: at the top of the Gafares sequence, directly below the transgressive marine Pliocene.

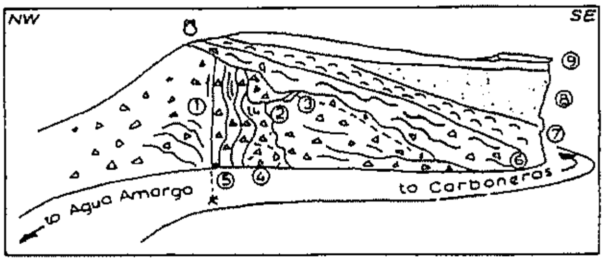

Fig. 4. Field-sketch of locality 8.
1 = Agua Amarga Breccia; 2 = subaerial scree of Agua Amarga Breccia material with (3) bleached zone at the top and (4) red clay fissure-fill; 5 = N.N.E. striking subvertical fault zone between Agua Amarga Breccia and downfaulted Feos rocks (2-4); 6-8 = marine Pliocene (Molata Formation) with (6) basal breccia with components derived from the Agua Amarga Member and *Ostrea* sp., (7) shell beds and (8) yellow sandstone; 9 = caliche crust.

Reddish lime-muds, that often contain irregulmar knobbly laminated levels, locally occur on top of the Agua Amarga Breccia around locality 1. Another example of Feos rock has been found in the nearby locality 8, in the last bend of the road to Carboneras, where it climbs out of Agua Amarga (Fig. 4). In the northern part of the basin (localities 3, 6 and around locality 9), some meters of pinkish-grey calcarenite and limestone conglomerate are present below the marine Pliocene. These rocks often occur in beds of up to several decimeters thickness, that are sometimes graded. Their components are comparable to those of the Agua Amarga Breccia, except for their better degree of rounding.

B. Molata Formation

This unit comprises a variety of clastic rocks, rich in marine fossils. It has been named after the conspicuous Molata hill, 6 kilometers to the northwest of Carboneras (see map, ca. 37° 01' N, 1° 57' W). The Molata Formation is transgressive upon the Gafares sequence or the volcanic basement of the Agua Amarga Basin.

Yellow to white bioclastic calcarenites, that can reach a thickness of several tens of meters, constitute the lower part of the Molata Formation in the northern-central part of the basin (e.g. localities 3, 4, 9, 10 and 11; see also Fig. 5). Some conglomerates containing well-rounded volcanic pebbles and fragments of Gafares sediments (e.g. Agua Amarga Breccia material), often occur at the base of the calcarenitic rocks, that are otherwise typically poor in terrigenous detritus. They are characteristically rich in bryozoan remains and further contain many benthic foraminifers and echinid spines and plate particles. Apart from these, fragments of molluscan shells and algae quite frequently occur. Planktonic foraminifers are more sporadic.

In the northern-central part of the basin, the calcarenites are covered by coarser-grained bioclastic limestones in which many algal remains and well-preserved shells of Pectinidae and especially *Ostrea* sp. are typical. Terrigenous material is found here as well-rounded matrix-supported pebbles. In the southern-central part of the basin, and on the higher parts of the irregular paleorelief of the Castillico Ridge, a few meters of the latter lithology builds the whole of the Molata sequence (e.g. localities 2, 5, 13, 6 and west of locality 10). This rests upon a remarkably flat and extensive abrasion surface that occurs around the present 100 meters topographic level around Agua Amarga.

A third facies type is found in the upper Molata rocks of the Castillico (loc. 12), and in those of a narrow strip to its S.S.E., stretching up to near Agua Amarga. Here, calcareous fossil-rich conglomerates and sands occur, showing an important influx of terrigenous material derived from the pre-Neogene alpine basement, exposed in the Sierras to the north of the adjoining Almeria-Nijar Basin.

Planktonic foraminifera are generally scarce, undiversified and poorly preserved in wash-residues from Molata sediments. Sometimes forms like *G. margaritae, G. puncticulata* and *G. crassaformis*, that are typical of the Mediterranean Pliocene have been encountered (cf. Addicott et al. 1978, who gave a short description of the bryozoan-rich calcarenites of our Molata Formation under 'Unnamed sandstone exposed to the south of Carboneras').

C. Campo de Nijar Formation

This unit typically consists of conglomerates and red silts. These have been first mentioned under this name by Ochoa et al. (1973b) from the central part of the Campo de Nijar, where they reach their major development. It is sometimes difficult to separate them from the lithologically similar (sub)recent deposits of the area. The Campo de Nijar rocks are however still somewhat tectonized.

Thin series more typical of this Formation have been found in the southwestern part of the basin (around locality 13). A few meters of caliche crust or calcified scree-breccia, that lies on top of the Molata Formation at a present topographic level above 100 meters could also belong to this unit.

FACIES INTERPRETATION AND DISCUSSION ON THE GEOLOGIC HISTORY OF THE BASIN AROUND THE MIO-PLIOCENE BOUNDARY

Open-marine and transgressive conditions during the earliest part of the Messinian are reflected by the richness of the marine fauna and fining-upward tendency in the lower part of the Turre Formation. Maximum depositional depths of 200-300 meters were probably created by this transgression. This has been firstly deduced from the fact that the most elevated, shallow marine Turre deposits occur at present topographic heights between two and threehunderd meters along the basin margins on SW-NE striking profiles across coastal areas of the basin (see also map). These have been hardly affected by differential uplift. The lower Abad marls from the deepest part of the basin, presently at topographic heights near sea-level, contain benthic foraminifer assemblages, the preliminary investigation of which by J.A. Manuputty is consistent with such waterdepth (pers. comm.). Our estimates further accord with the minimum figures suggested by Roux & Montenat (1977) for the Messinian of the area, which was based on echinoïd remains from rocks exposed near Cortijo El Plomo (cf. our locality 2), and in all probability derived from the Turre Formation in one of the deepest parts of the basin.

The rather strong increase in influx of reef-derived material in the upper Turre rocks of more central parts of the basin, suggests the onset of regressive conditions towards the end of the early Messinian. This impression is further strengthened by the preliminary investigation of the benthic foraminifer assemblages from the younger Abad marls, indicating deposition under shallower waterdepths. The same general trend is reflected by the progradation of the marginal *Porites* reefs over their former reef slopes, which has been described by Dabrio & Martin (1978) and Esteban & Giner (1980) from the Agua Amarga area, and by inter alia Völk (1967) and Pagnier (1977) from the nearby Vera and Sorbas basins.

The oligotypical foraminifer assemblages from the upper part of the Abad Member attest to restriction of the environment contemporaneous with the regressive conditions. The restriction became more or less complete during the subsequent deposition of Manco sediments. Their virtual lack in fossils and the presence of small cavities in the form of gypsum crystals point to evaporitic circumstances.

We believe that the facies interpretation given in this paragraph contains some indications for the setting in which the Agua Amarga Breccia originated. Hypotheses regarding this event are separately discussed below. The hypothesis of a formation by means of solution-collapse is favoured.

Due to their facies and stratigraphic position, the succeeding Feos deposits are clearly related to an important phase of erosion and karstification that occurred at the end of the Gafares depositional cycle. The reddish limestone occurrences at the top of the Gafares sequences of localities 1 and 8 are interpreted as caliche soil and scree formation. Metre-wide karstic hollows have been found in the top of the Agua Amarga Breccia of locality 14 (Fig. 6). The subrounded flat-pebble conglomerates, that occur locally in the northern-central part of the basin, attest to reworking of Agua Amarga Breccia material. These rocks may well represent fluviatile deposition. Evidence for a rounded palaeorelief on the surface of the Agua Amarga Breccia can be observed in the left bank of the northern branch of the Barranco del Hondo (locality 11). The Breccia here rather abruptly thins out from a thickness of 20, to 0 meters, over a distance of 200 meters (see also Fig. 5). Further good evidence for such a palaeorelief was found in the upper reaches of the Rambla del Cinto (locality 10). In between the latter localities, a relatively wide valley was excavated, whose relief was probably somewat smoothed-off during the subsequent Pliocene transgression. The erosion that created this valley has not only cut into the Agua

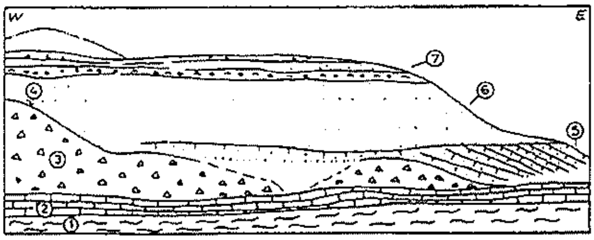

Fig. 5. Field-sketch of locality 11 (Barranco del Hondo). 1 =marls of the Abad Member (Turre Fm.); 2 = reef-derived debris-flows of the Cantera Mb. (Turre Fm.); 3 = Agua Amarga Breccia; 4 = erosional surface; 5-7 = Molata Formation (marine Pliocene): 5. mega-crossbedded calcarenite; 6 = greenish-yellow sands, and 7. shell-rich conglomeratic sandstone

Amarga Breccia. In a number of localities in the Rambla Oliveira and in the lower reaches of the Rambla del Hondo, part, or all of the underlying Turre sequence has been removed and the Pliocene Molata rocks are deposited directly upon the volcanic basement of the basin.

All these phenomena attest to a period of maximum regression at the end of the Gafares depositional cycle, which has resulted in a widespread unconformity between the Messinian and Pliocene sediments. Its presence has previously been recorded from localities around the Mesa Roldan by Montenat et al. (1976), Addicott et al. (1978) and Esteban and Giner (1980). The fall in sea-level in the Agua Amarga Basin since the maximum transgression in the early Messinian is estimated to have been about 250 meters. This has been calculated from the topographic height diffference between the most elevated, marginal early Messinian deposits and the deepest point of incision of the pre-marine-Pliocene erosion surface (again on SW-NE striking profiles across coastal areas of the basin; see also map).

The regression is considered to have been partly due to tectonic uplift of the studied area. The presence of tectonic activity in the Mio-Pliocene transition follows most directly from the somewhat stronger tectonization of the Messinian beds, compared to the overlying marine Pliocene. Especially near the Castillico Ridge, the former often play somewhat higher dips. Bousquet & Philip (1976) have described the presence of a fracture system in the Messinian limestones of the northern edge of the Playa de los Muertos (our locality 4), that is not present in the overlying marine Pliocene. The small step-fault that effects the Gafares rocks of locality 11 (Fig. 4), may well represent the same phenomenon. The maximum waterdepth of the early Messinian sea in the Agua Amarga Basin appears to have been more considerable than that calculated for sedimentation during the younger, Pliocene transgression (see below). Their depositional depth difference is estimated at about 150 meters, which cannot be matched by sedimentary infill in the meantime. A net tectonic upliftof the basin of such an amount is thus likely to have taken place between the two transgressive periods.

It is however considered probable, that the widespread lowering of sealevel at the end of the Miocene, of which inter alia Hsü et al. (1973, 1977), Van Couvering et al. (1976) and Adams et al. (1977) have described effects from areas both within and outside the Mediterranean realm, also had an influence on the regressive and erosive circumstances towards the end of the Gafares sedimentation.

Newly transgressive and open-marine conditions are reflected by the fining-upward tendency and richness in marine fauna of the older Molata sediments. The maximum waterdepth during the Pliocene Molata deposition is estimated at 100-150 meters. This is firstly suggested by the general neritic character of the bryozoan-rich calcarenites of the deepest, northern-central part of the basin. In these, benthic foraminifers often dominate over their planktonic equivalents. A relatively shallow depth of deposition for these sediments can further be indirectly deduced from their geometric relation to the thin marginal sequences, that constitute the whole of the Molata series in relatively higher parts of the basin, presently around the 100 meters contour (their presently higher position on the Castillico Ridge is ascribed to younger, SE directed low-angle tilting).

The shallow marine character of the latter sediments follows on the one hand from the *Ostrea* -conglomerate facies, in which they are often found. The calcareous sands and conglomerates on the other hand, that occur in locality 14 on a similar topographic height, contain foraminifer assemblages in which *Cibicides lobatulus, Elphidium* sp. and *Ammonia beccarii* strongly predominate, thus indicating depositional depths less than about 50 meters here. The present escarpments of the Rellana de San Pedro, can thus be envisaged as a steep cliff-coast along a shallow bay in Pliocene times.

THE AGUA AMARGA BRECCIA: A COLLAPSED LIMESTONE DEPOSIT?

The question of the origin of the Agua Amarga Breccia is still open to discussion. What mechanism can have produced this chaotic limestone mass of several tens of square kilometers areal distribution and a thickness of up to 40 meters.

The present deposits show a striking similarity on several points, both on basis of description (West 1975) and personal observation by one of us (Th. B.R.), to the so-called 'Broken Beds' from the Purbeck (Upper Jurassic) of the south coast of England. These are however generally interpreted as the product of tectonic brecciation (e.g. West in op. Cit.), a mechanism unlikely to have been the major cause of formation of the Agua Amarga Breccia, which is mostly tilted over a few degrees only, in addition to a few small scale step faults (like the one described from locality 8).

Another example of strongly similar rock is the Carboniferous '*Cyatophyllum* Limestone breccia' of West Spitsbergen. It has been described and interpreted by McWhae (1953) as the result of collapse after dissolution of underlying evaporites. Such a possible mode of origin for the Agua Amarga Breccia will be further discussed below.

Still another mechanism could be deposition by means of mass gravity transport. The absence of bedding in the Agua Amarga Breccia and its composition of angular fragments of strongly varying grainsize would suggest deposition, either as a subaerial scree or as part of a slumped mass or giant debris flow. The first alternative is supported by the nature of the upper surface of the breccia, where soils, karstic surfaces and widespread erosion occur. It is however contradicted by the regionally smooth lower surface and vast, sheetlike aereal distribution. In this regard it should be mentioned, that the lateral thinning and wedging out of this unit towards the Castillico Ridge and Rambla Oliveira

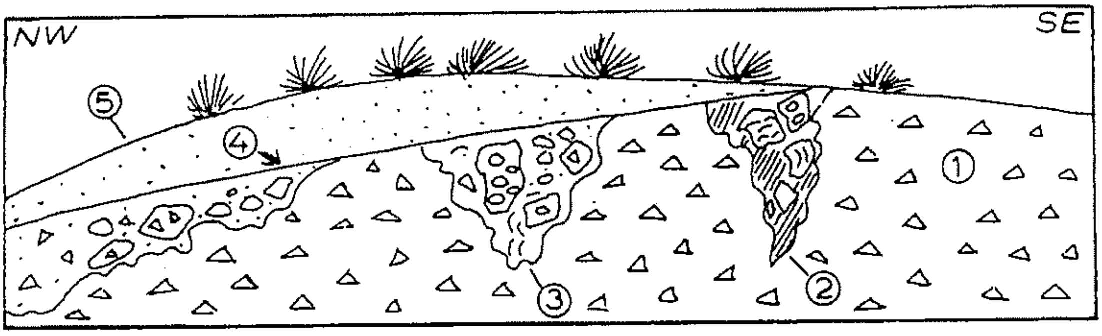

Fig. 6. Field sketch of locality 14. 1 = Agua Amarga Breccia. 2 = Karstic fissure with red earth and limestone blocks containing molluscan shells and well-rounded volcanic pebbles. 3 = As 2, filling consists of limestone blocks in a matrix of Pliocene material. 4 = Abrasion surface. 5 = Greenish-yellow fossiliferous sands of the Molata Formation.

region is, to at least a large extent, due to post-depositional erosion prior to the Pliocene transgression, since this erosion likewise affects the underlying Turre deposits in these areas. The nature of the constituent material of the breccia further conflicts with a subaerial scree origin: the highest nearby (paleo) relief element (Mesa Roldán and Rellána de San Pedro) consist of volcanic basement with a thin cover of the marginal Messinian *Porites* reef-bearing series; no volcanic material and only a few fossiliferous limestone fragments have been found in the breccia.

The features of the Agua Amarga Breccia are thus best explained, either by deposition from a giant debris-flow, or by means of dissolution-collapse.

In the case of breccia formation by slumping and debris-flow, both the amount of material involved in the breccia and its uniformity imply deposition through one 'catastrophic' event. This possibility further requires the original presence of an unstable limestone mass in a topographically high position and a presently incomplete stratigraphy in the source area. The Castillico Ridge and the area to its north, northwest of Carboneras, would, at first sight, seem to satisfy such conditions (see map). However, the marine Pliocene Molata deposits at the southern flank of the Ridge are tilted towards the centre of the Agua Amarga basin and develop no fundamental difference in facies from the Castillico area, where they have retained a horizontal position, untill several tens of meters lower in the Agua Amarga Basin. The present elevation of the Ridge above the topographic level of the Agua Amarga Breccia (ca. 100 meters on average) is therefore largely due to relatively recent differential uplift. It is thus questionable whethet a steep enough slope existed between this area and the centre of the Agua Amarga basin.

An alternative source for the breccia could be somewhat further away: the area of the southwest-northeast trending 'Serrata-Carboneras Fault Zone', which dissects the eastern part of the adjoining Nijar Basin (part of its is visble on the map, Fig. 1). Recent sinistral strike-slip movement has taken place along the faults, as shown by Quaternary offset of river courses, man-made structures in the Campo de Nijar and the seafloor to the southwest in the Gulf of Almeria (Greene et al. 1977). Tectonic movements in this zone may have already begun in Early Neogene times (e.g. Bousquet et al. 1975, Armijo et al. 1977). An elongate positive relief with culminations of over 300 meters presently marks the area of the fault zone. That part of it directly to the west of the Agua Amarga Basin (Molata-Caballón region, see map), is an unlikely source-area for the breccia since practically complete Messinian sequences with gypsum are here often found, whereas the marine Pliocene occurs in a relatively deep marine facies. However, the Agua Amarga Breccia material may have been conceivably derived from an area subject to fault zone tectonics, which, by sinistral tanscurrent movements, was later displaced to the southwest relative to the Agua Amarga basin.

Finally, the possibility of brecciation as the result of collapse of brittle limestone beds after dissolution of directly underlying or intercalated salt layers remains to be discussed. The virtual absence of fossils and the presence of limestones with small cavities with the shape of gypsum crystals suggest original deposition of both the Agua Amarga Breccia material and the locally urnderlying, thin Manco series in an bnormal marine and, at least from time to time, evaporitic environment. The occurrence of totally calcified selenitic gypsum lenses in the upper part of the Gafares sequence of locality 3 (Rambla del Pozo), points to the original presence of more considerable amounts of Messinian evaporites in the deeper parts of the Agua Amarga Basin (Dronkert & Van de Poel, in prep.). A further argument stems from the regional stratigraphic setting. The Agua Amarga Breccia occurs in central parts of the basin, on top of the youngest Turre deposits and in close relation to the 'Calcare di Base-like' Manco rocks; it thus occurs in the same stratigraphic position as the Messinian gypsum in the adjoining Almeria-Nijar area, and the nearby Sorbas Basin (see map and compare Dronkert 1976/1978, Dronkert et al. 1979).

A second prerequisite for this hypothesis is the subsequent presence in the basin of a considerable amount of water of low salinity to effect the leaching out of the more soluble salts deposited earlier. Evidence has been found for the presence of continental conditions after (or coïncident with?) the formation of the breccia, prior to the deposition of marine Pliocene. The local presence of a valley-like morphology and the facies of some sands and conglomerates of the Feos Member indicate fluviatile conditions. On a regional scale, Roep & Van Harten (1979) have deduced that an appreciable amount of fresh water gained access to a numebr of the Neogene basins in southeastern Spain at the end of the deposition of the Messinian evaporites. It may be remarked that the sediments from the top of the Messinian of several basins of this area show affinity, both in their facies and their stratigraphic position, to the typical 'lago mare' deposits from more eastern parts of the Mediterranean area as described by i.a. Ruggieri & Sprovieri (1976) and Vismara Schilling et al. (1978).

Although we do not completely exclude the posssibility of a mass-flow origin for the Agua Amarga Breccia at present, we think that our data are most easily explained by the solution-collapse hypothesis. We thus conclude that a collapse of limestone layers, due to the dissolution of underlying or intercalated salts, has most likely been the mechanism for the formation of the breccia.

CHAPTER 5

SYNTHESIS AND DISCUSSION

Mio-Pliocene lithostratigraphic development in the Eastern Almeria Province is relatively uniform. The most important difference exists in the better development of the main middle and late Messinian units in the inland Sorbas and Northern Nijar basins compared to the coastal Agua Amarga and Vera basins, whereas the opposite is true for the marine Pliocene series (Fig. 5.1).

Primarily based on the lithologic, microfossil and authigenic mineral development in central basin sections, a number of distinct, subsequent integrated facies units can be recognized in the Mio-Pliocene time-interval (Fig. 5.2, right). The characteristic component assemblage of each facies indicates its environmental setting (Fig. 5.2, topline), in terms defined in Chapter 1 (Fig. 1.2). A summary of their interpretation in physico-chemical environmental parameters is given in Fig. 5.3. Successive phases with relatively uniform environmental conditions are distinguished as 'stages' (Fig. 5.3, right). They were defined in the Northern-Nijar Basin (Ch. 3), but the similar vertical facies development in the other basins of the EAP shows a basically uniform development of environmental conditions over the entire area (Ch. 4; Figs 5.1-5.3).

Major vertical facies changes, often correlated with unconformities in marginal sections, represent changes in environmental conditions which I numbered as successive 'events' (Figs 5.1-5.9). Our main Messinian facies units compare with those of the 'standard stratigraphy' from the central Mediterranean and the 'event-stratigraphic scheme' of Müller & Hsü (1987) for that area has been followed as closely as possible in the numbering of our events (Figs 5.4 & 5.5).

The age of local environmental stages and events can, more or less precisely, be determined by means of planktonic foraminifer biostratigraphy, calculations on basis of sediment accumulation rates, and to some extent by comparison with well-dated facies equivalents from within the same region.

Combination of our data with those from Central Mediterranean sections, mostly derived from the literature, gives information on the regional setting in which deposition took place and gives insight in the underlying factors of the facies development (Fig. 5.4-5.6). Evaluation of the paleoenvironmental development in the EAP on the other hand adds new data for the interpretation of the paleoceanographic history of the Mediterranean Basin and the pattern of water-exchange with the open ocean through its Entrance Area (Figs 5.5 & 5.6).

Investigations into the stratigraphic organization of the facies, besides revealing their lateral extension, emphasize their vertical and lateral changes and the presence or absence of unconformities. A summary of our interpretation of main aspects of the stratigraphic development in the EAP is given in Fig. 5.4. Sequence stratigraphy and more classical forms of basin analysis (Figs 5.7-5.9) further add to the understanding of the local basin development and further address the underlying factors of environmental change (climatic and tectonic geodynamic processes).

SUMMARY OF FACIES DEVELOPMENT AND ITS INTERPRETATION IN TERMS OF LOCAL AND REGIONAL ENVIRONMENTAL SETTINGS

In all the studied sections, the Mio-Pliocene transition is first characterized by common lamination, low marine and continental fossil diversity, extreme variations in fossil abundancy, high authigenic mineral content, and common organization in well-developed cyclic sequences as the result of variations in interrelated main physico-chemical environmental parameters (waterdepth, temperature, salinity, oxygen and nutrient content (Figs 5.1-5.3). These features are characteristic of restricted marine (marginal-basin), and restricted continental sedimentation in relatively deep water (Ch. 1).

Towards its top, these features are associated with coarse terrigenous-clastic deposits, subaerial condensation horizons, karstification and/or erosional hiatuses, which are primarily characteristic of the marginal position of our area, at least towards the end of the Messinian. These Messinian-earliest Pliocene strata are intercalated between open-marine Tortonian and younger Pliocene deposits, in which marginal basin characteristics are virtually lacking. In all the basinal areas maximum depositional depths during the Pliocene were distinctly smaller than during the earliest Messinian (Figs 5.3, 5.7-5.9).

FIGURE 5.1 Schematic lithostratigraphic columns for the Mio-Pliocene of the central parts of the basins of the EAP and their proposed correlation (Chs. 3 & 4). Bold lines between columns mark boundaries ('events') between major facies units (I-V) representing 'Environmental stages'. They are interrupted where diachroneity has been established or where synchroneity is not known. Thicknesses and lithostratigraphic terminology for the Sorbas and Vera basins from Dronkert & Pagnier (1977) and Völk & Rondeel (1964; with additional distinction of Gypsum breccia and *Cyprideis* marl proposed here).

Note the basic uniformity in facies development between the different basinal areas. Differences in development of the major facies units appear in the middle-upper Messinian and Pliocene and are ascribed to differences in geodynamic and paleogeographic setting of the basins: uplift for the Agua Amarga, lack in subsidence and sedimentary infill for the Sorbas, and relative subsidence for the Northern Nijar and Vera basins, major erosion during late Messinian Mediterranean drawdown in the more open, coastal Vera and Agua Amarga basins.

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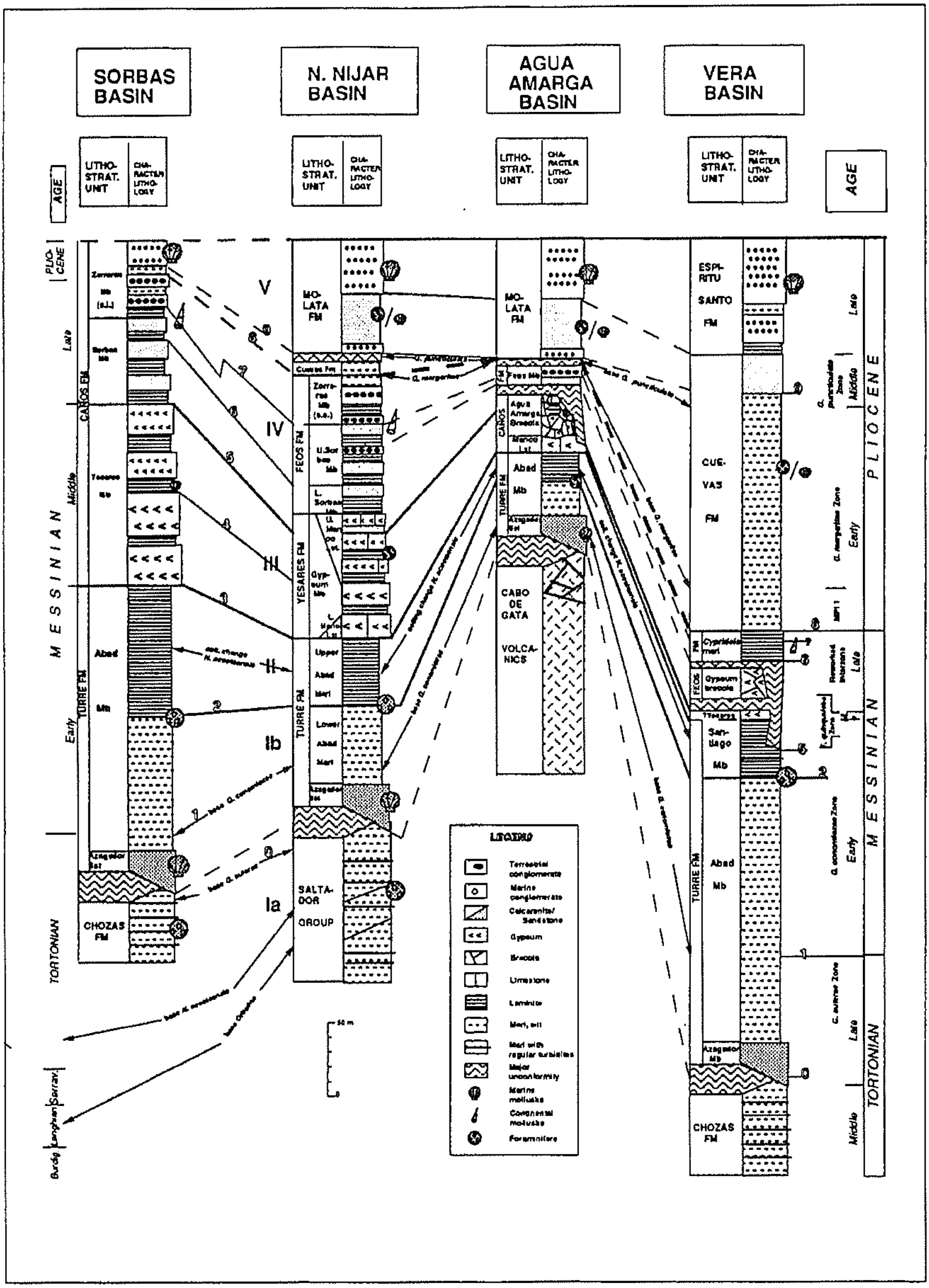

SORBAS BASIN
N. NIJAR BASIN
AGUA AMARGA BASIN
VERA BASIN
AGE
LITHO-STRAT. UNIT
CHA-RACTER LITHO-LOGY
PLIO-CENE
MESSINIAN
TORTONIAN
PLIOCENE
Late
Middle
Early
CAÑOS FM
TURRE FM
Abad Mb
Azagador Mb
CHOZAS FM
MO-LATA FM
FEOS FM
YESARES FM
Upper Abad Marl
Lower Abad Marl
SALTA-DOR GROUP
CABO DE GATA VOLCA-NICS
ESPI-RITU SANTO FM
CUE-VAS FM
San-tiago Mb
V
IV
III
II
Ib
Ia
50 m
0
LEGEND
Terrestrial conglomerate
Marine conglomerate
Calcarenite/ Sandstone
Gypsum
Breccia
Limestone
Laminite
Marl, silt
Marl with regular turbidites
Major unconformity
Marine mollusks
Continental mollusks
Foraminifera

Pre-Messinian Neogene conditions in the EAP are represented by the base of the Turre Formation and the underlying Tortonian and 'Older Neogene' stratigraphic units ('Saltador Group'; Fig. 5.1). Burdigalian to Tortonian age is indicated by the presence of N5-base N17 planktonic foraminifer zones (Chs. 3 & 4.3; Völk 1967, Montenat et al. 1976, Gonzalez Donoso & Serrano 1977, unpublished data of L.P.A. Geerlings, J.A. Manuputty and the author). The relative development of these units decreases with their age and exposure of the 'Older Neogene' series is essentially restricted to major fault zones of the area (Ch. 2).

Blue-grey, massive '*Globigerina marl*', with diverse planktonic and benthic foraminifers and common glauconite is the dominant lithology, which, together with a number of other characteristic components is typical of 'Facies I' (Fig. 5.2). Higher benthic foraminifer diversity in respect to the overlying earliest Messinian and common turbiditic intercalations allow for the recognition of a *Subfacies Ia* for the interval under discussion. Important episodes with restricted marine and even continental facies components have been recognized in the Vera, Sorbas and Nijar basins (Rondeel 1965, Völk 1967, Brunsman 1981; unpublished data). These are probably of more than local importance in particular as a late Serravallian episode is concerned (Dronkert et al. 1979). However, the dominance and consistent recurrence of typical Facies Ia is here retained as most characteristic for the pre-Messinian.

Benthic faunas indicate common deposition in marine basins without major oxygen deficiency or salinity deviations under waterdepths of many hundreds of meters ('mesobathyal') (Ott D'Estevou 1980, Kleverlaan 1989; Fig. 5.3). Completely hispid uvigerinas, *Oridorsalis* spp. and *Cibicidoides italicus* and *C. wuellerstorfi* are relatively common in the benthic foraminifer assemblage ('Subassemblage A_0'') (Völk 1967; unpublished data, Fig. 5.2). These forms are presently rare or absent in the Mediterranean, but characterize modern deep ocean environments with cold, 'psychrospheric' bottom waters (Wright 1978a & b, Hermelin & Scott 1985, Lutze 1986, Van Marle 1988, Hasegawa et al. 1990, Gupta & Srinivasan 1992). Surface waters were warm subtropical as attested by the planktonic foraminifers and carbonate platform biota.

Burdigalian-Tortonian deposits with equivalent facies characteristics are common in more central parts of the Mediterranean (Benson 1973, 1978, Ricci Lucchi 1973, Wright 1978a & b, Colalongo et al. 1979, Meulenkamp et al. 1979, Dondi & Rizzini 1979/1980, Van der Zwaan 1982; Fig. 5.4). They further attest to a predominant 'deep, open oceanic' regional setting for this period (Figs 5.5 & 5.6). Regular siliceous intercalations with common radiolarians in Burdigalian-Tortonian deposits of our area and other basins of the Betic Cordilleras, probably represent coastal upwelling of a relatively open-marine type in a periodic 'restricted ocean margin' setting (Völk 1967, Hermes & Smit 1977; unpublished data of the author; Fig. 5.2.Ia).

The late Tortonian, where data on the deeper-water development are relatively scarce because of the presence of an important regressive episode with widespread erosion and shallow-water deposition of the Azagador Sandstone (*'Event 0'*) in S.E. Spain, probably represents a transitional interval with a first tendency to marginal basin conditions. Well-developed open-marine, warm-water conditions in the higher part of the watercolumn are still indicated by the micro and macrofauna constituents (Subassemblages A_0 and A_1) of our area.

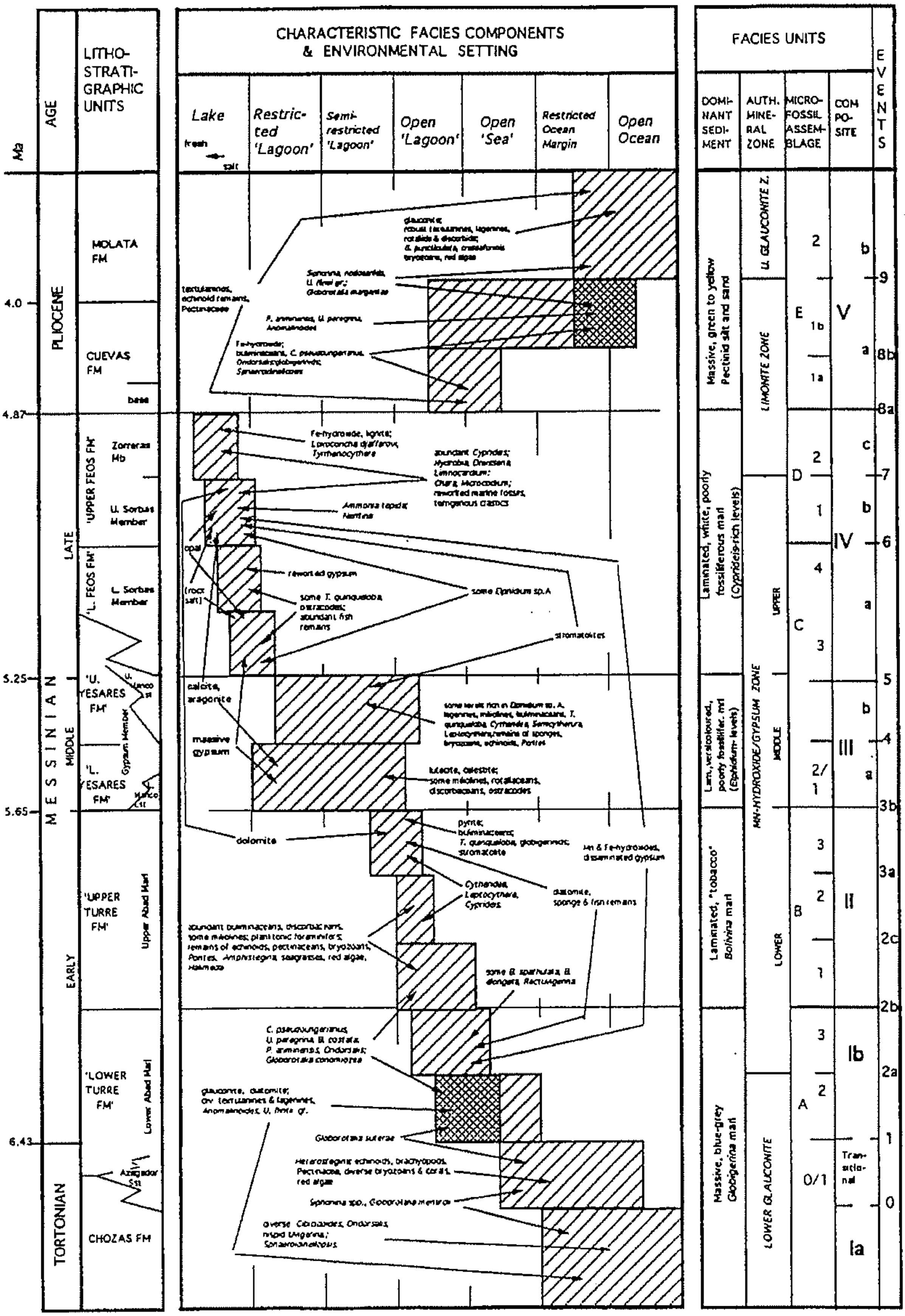
Data on deeper water sections in more central parts of the Mediterranean still show the presence of some 'open oceanic' benthos (e.g. hispid uvigerinas and some diverse *Cibicidoides*), but at the same time suggest the disappearance of the planktonic *Sphaeroidinellopsis* and *Globoquadrina* and a decrease in abundance of typical *G. menardii*, whereas the appearance of the 'endemic' *Globorotalia suterae* is a common event in Mediterranean sections in this interval (cf. Zachariasse 1975, 1979, Colalongo et al. 1979, Van Hinte et al. 1980). In particular this planktonic development differs from that recorded in external parts of the Mediterranean Entrance Area (Ch. 3.3, Benson et al. 1991).

The earliest Messinian is represented by the main part of the 'Lower Abad Marl', which contains the main part of the *G. conomiozea* planktonic foraminifer Zone and is well-developed all over the investigated area (Fig. 5.1).

Massive, blue-grey '*Globigerina*-marl' with abundant, relatively diverse planktonic and benthic foraminifers is still the dominant lithology of this interval. Some reduction in benthic foraminifer diversity (disappearance of *Siphonina*), the replacement of *Globorotalia menardii* by *G. conomiozea*, the virtual disappearance of turbidite intercalations and an increase in frequency of regular silicified diatomitic beds define *"Event 1"* and differentiate '*Subfacies Ib*' from that of the underlying interval (Figs 5.1, 5.2, 5.7).

The characteristic facies components still indicate deposition in a relatively deep and open-marine setting. Major oxygen-deficiency, probably accompanied by some increase in salinity and bottom-water temperature and some shoaling are not recorded before the uppermost part of 'Substage Ib', but the entire interval may represent somewhat cooler climatic conditions than in the preceeding Tortonian (Figs 5.2 & 5.3; Ch. 4.3, Saint-Martin & Rouchy 1990).

Characteristic component composition in our area and comparison with time-equivalent sections from the open, western end of the Mediterranean Entrance Area, first indicate slight restriction of the marine environment and predominant conditions which are in many


Ma
AGE
LITHO-STRATI-GRAPHIC UNITS
CHARACTERISTIC FACIES COMPONENTS & ENVIRONMENTAL SETTING
Lake
fresh
salt
Restric-ted 'Lagoon'
Semi-restricted 'Lagoon'
Open 'Lagoon'
Open 'Sea'
Restricted Ocean Margin
Open Ocean
FACIES UNITS
DOMI-NANT SEDI-MENT
AUTH. MINE-RAL ZONE
MICRO-FOSSIL ASSEM-BLAGE
COM PO-SITE
EVENTS
4.0
4.87
5.25
5.65
6.43
PLIOCENE
MESSINIAN
LATE
MIDDLE
EARLY
TORTONIAN
MOLATA FM
CUEVAS FM
base
Zorreras Mb
'UPPER FEOS FM'
U. Sorbas Member
'L. FEOS FM'
L. Sorbas Member
'U. YESARES FM'
'L. YESARES FM'
Gypsum Member
'UPPER TURRE FM'
Upper Abad Marl
'LOWER TURRE FM'
Lower Abad Marl
Azagador Sst
CHOZAS FM
calcite, aragonite
massive gypsum
dolomite
opal
(rock salt)
reworked gypsum
stromatolites
Massive, green to yellow Pectinid silt and sand
Laminated, white, poorly fossiliferous marl (Cyprideis-rich levels)
Massive, blue-grey Globigerina marl
Laminated, "tobacco" Bolivina marl
U. GLAUCONITE Z.
LIMONITE ZONE
MN-HYDROXIDE/GYPSUM ZONE
UPPER
MIDDLE
LOWER
LOWER GLAUCONITE
E
D
C
B
A
2
1b
1a
4
3
2/1
0/1
V
IV
III
II
Ib
Ia
Tran-sitio-nal
a
b
c
9
8b
8a
7
6
5
4
3b
3a
2c
2b
2a
1
0

aspects comparable with the present 'open' Mediterranean Sea (Figs 5.2 & 1.2). Facies development in the EAP is comparable to that in other marginal sections of the Mediterranean, but in its deepest parts a high degree of restriction of the bottom waters is already recorded under a relatively normal marine upper water layer of at least a few hundred meters thickness (Ch. 3.3).

Some 'aberrant' features of these upper waters (absence of *Siphonina*, presence of keeled globorotaliids and *Oridorsalis*), probably indicate a relatively high nutrient level. High nutrient content has been recorded in laminated interbeds in different parts of the Mediterranean, where they are associated with periodic salinity increase and/or cooling (Bizon et al. 1979, Van der Zwaan 1982). Diatomitic interbeds in the Agua Amarga Basin and regular siliceous intercalations recorded in other areas of the EAP (Chs. 3 & 4), probably represent the same phenomenon. A relation with strong inflow of Atlantic water during periods of relatively high evaporation over the Mediterranean was proposed by Van der Zwaan (1982), whereas Benson et al. (1991) suggested that such water should be nutrient-rich due to the 'siphon-effect'. Episodic 'inwelling' of nutrient rich intermediate ocean-water is observed at present and during Pleistocene 'glacial' periods in the Red Sea Basin (Ganssen & Kroon 1991). Our data, combined with those from other parts of the Mediterranean suggest an earliest Messinian regional setting of the EAP at the upper slope of a marginal basin with still relatively minor restriction, primarily due to common, periodic 'inwelling' and related nutrient-excess and oxygen-deficiency (Figs 5.3, 5.5 & 5.6).

The late early Messinian is represented by the 'Upper Turre formation', which comprises the Upper Abad Marl, the local Santiago Turbidite Member, and the common Cantera Reef Limestone of the margins of the basins of the EAP (Figs 5.1 & 5.7). This interval is well-represented throughout the basinal areas with a slightly lesser development in the

<----- FIGURE 5.2. Summary of facies characteristics in a composite diagram of the lithostratigraphic units of the EAP and their interpretation in terms of confinement degree of the marine environment (topline). Characteristic facies components as described in Chapters 3 & 4 with some additional data from Völk 1967, Montenat et al. 1976, 1980, Roep & Van Harten 1979, Ott D'Estevou 1980, Dronkert 1985, Van de Poel & Andreu unpublished. Tortonian and early Pliocene essentially from the Vera Basin, Messinian and middle Pliocene largely from the Northern Nijar Basin with additional data from the Sorbas and Agua Amarga basins (cf. Fig. 5.1). No vertical scale.

Arrows mark lower and upper limits of facies components. Hatched areas denote range of environmental setting indicated by the characteristic facies components from the EAP. Range of environmental settings for the EAP can sometimes be further limited (dark areas) by considering stable isotope data and/or facies composition from Central Mediterranean sections.

Note the gradual increase in restriction of the area at the end of the Miocene and the sudden, two-step, return of open marine conditions at the beginning of the Pliocene.

Vera Basin (Chs. 3 & 4). Late early Messinian age is indicated by the presence of top and base of the Mediterranean *G. conomiozea* and *T. quinqueloba* planktonic foraminifer zones, respectively, and calculations on basis of sedimentation rates.

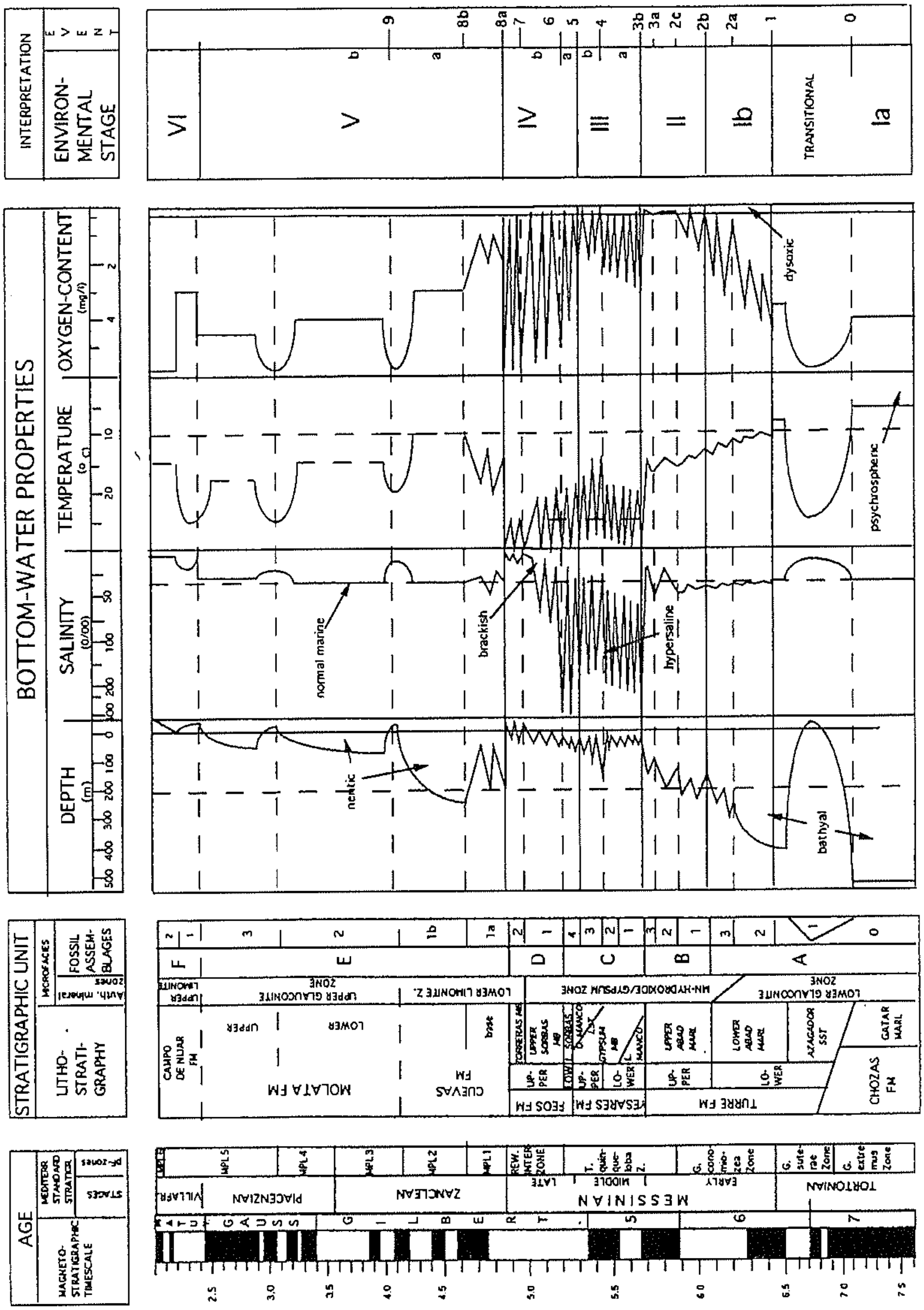
Yellowish ('tobacco-coloured'), laminated '*Bolivina* marl' is the dominant lithology of 'Facies II', which is further characterized by a microfossil assemblage with relatively low diversity and high abundancy and a moderately high authigenic mineral content (common disseminated small gypsum crystals and iron and manganese hydroxides, some pyrite and dolomite and a few concentration levels of these minerals) (Fig. 5.2). The facies changes recorded near the base of this interval represent *'Event 2'*. It marks a gradual evolution, since features characteristic of Facies II periodically appear in the top of Stage Ib, whereas they become exclusive in the middle-upper part of this interval only (Ch. 3.3; Figs 5.2 & 5.3).

Predominant oxygen-deficiency at the sea-bottom, at least periodic nutrient excess, common, but still relatively weak excursions into the hyper and hyposaline field of the salinities of bottom and surface waters, and outer neritic depositional depths characterized the basin centres of the EAP during 'Environmental stage II' (Fig. 5.3). Common, cyclic, interrelated fluctuations of the different environmental parameters are recorded with a gradual increase in restriction and some upwards shoaling during this interval. Surface-water-temperatures near the outer limits of a subtropical region (18^{o} C in winter) are suggested by diverse data on fossil assemblages (Ch. 4.3, Esteban 1980, Saint-Martin 1987, Jimenez & Braga 1993).

The characteristic components of our Facies II compare with those of

FIGURE 5.3. Evolution of Mio-Pliocene bottom-water properties in the deeper parts of the EAP and its interpretation in terms of environmental stages and events. Bottom-water properties are mainly based on interpretation of the Mio-Pliocene facies in the central part of the Northern Nijar Basin (Chs 3.2 & 3.3), with additional data for the Messinian from the Sorbas and Agua Amarga basins, and interpretation of the Tortonian 'Chozas Formation' and early Pliocene 'Cuevas Formation' in Sorbas and Vera basin sections (compare Fig. 5.1). Values for bottom-water properties are approximate and are primarily meant to illustrate relative changes. Periods with relatively uniform parameter characteristics are numbered as environmental (sub)stages, whereas their boundaries ('events') mark changes in environmental setting.

In particular the middle and late Messinian shows typical marginal and continental basin characteristics (strongly aberrant salinities and water temperatures). Oxygen-deficiency and a strong cyclic development are other features of the Messinian period. Common shallow waterdepths are partially the effect of periodic and episodic drawdown of global and/or Mediterranean 'sea' level, partially of a general shoaling tendency in the Mio-Pliocene interval due to local tectonic and sedimentary processes (uplift and infill) (Figs 5.7-5.9) ------------------>


AGE
MAGNETO-STRATIGRAPHIC TIMESCALE
STRATIGRAPHIC UNIT
LITHO-STRATI-GRAPHY
FOSSIL ASSEM-BLAGES
BOTTOM-WATER PROPERTIES
DEPTH (m)
SALINITY (0/00)
TEMPERATURE (° C)
OXYGEN-CONTENT (mg/l)
INTERPRETATION
ENVIRON-MENTAL STAGE
EVENT
neritic
bathyal
normal marine
brackish
hypersaline
psychrospheric
dysoxic
TRANSITIONAL
MOLATA FM
CUEVAS FM
TURRE FM
CHOZAS FM
TORTONIAN
MESSINIAN
ZANCLEAN
PIACENZIAN

the essentially time-equivalent Tripoli Formation and of many other late early Messinian sections from the central Mediterranean which attests to a Mediterranean Basin with relatively uniform conditions, separated from the open marine environment by a major sill somewhere in between our area and the basins in the outer part of the Gibraltar Arc (Figs 5.4 & 5.5; Chs. 3.3 & 4.3, Gersonde & Schrader 1984, Gautier et al. 1994). The low fossil diversity throughout the basin and the common 'weakly euryhaline' character of the fossil constituents, further indicate, together with stable isotope data, typical enclosed ('lagoonal') marginal basin conditions (Ch. 3.3, Rouchy 1982). The abundance in typical marine organisms as planktonic foraminifers, corals, marine mollusks, echinoids, red algae, still indicates a relatively good connection with the open marine environment, on the other hand (Sturani 1973, 1978, Montenat et al. 1980, Riding et al. 1991b, Ch. 3) The characteristic component assemblage of our area is comparable in a number of aspects with the Recent NW Adriatic, Kau Bay (Indonesia) or the Mediterranean and Red Sea Basins during Pleistocene episodes and a predominant 'open lagoonal' regional setting is proposed (Ch. 3.3; Figs 1.2, 5.5 & 5.6). The diatomite intercalations of this interval probably represent periods with an improved marine connection during relative highstand, which, in combination with increased run-off under relatively wet regional climate, may have led to periodic estuarine circulation (Müller & Schrader 1989; Fig. 5.6). Evidence for a link between humid Mediterranean climate and global interglacial conditions during the Pleistocene was presented by Suc & Zagwijn (1984) and Combarieu Nebout (1987, *et al.* 1990).

The **middle Messinian** is represented by the main part of the 'Yesares Formation' (Gypsum Member, Manco Limestone, Agua Amarga Breccia and marginal Oolite Member), which is best preserved in the Sorbas and Northern Nijar Basins (Figs 5.1 & 5.7). Its (?virtual) absence in the Vera Basin and its lesser development and strong alteration in the basin of Agua Amarga are primarily explained by late Messinian and early Pliocene erosion and karstification (Ch. 4).

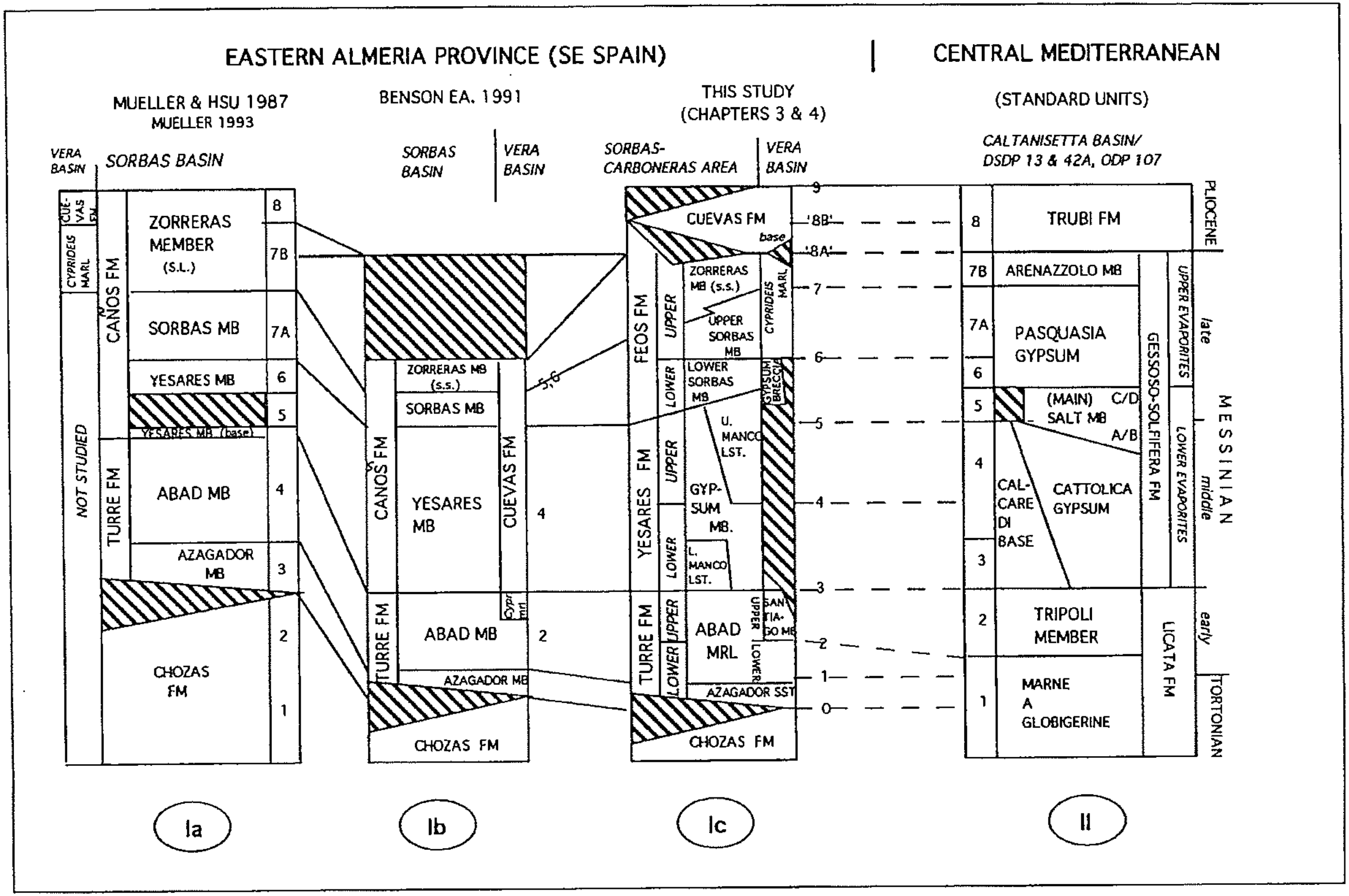
The middle Messinian age is primarily indicated by the conformable position upon early Messinian sediments in basinal sections, calculation of the age of the base of this interval on basis of sedimentation rates in the underlying Abad Marl and possible correlation of a marine interval in the upper part of our Yesares series with the Fortuna Basin of E. Spain (Ch. 3).

'Facies III' characterizes this interval, where extremely *poorly fossiliferous, versicoloured laminated marl* with thick gypsum intercalations is the predominant lithology (Figs 5.1 & 5.2). It is further primarily characterized by its richness in other authigenic minerals (dolomite, calcite, aragonite, sedimentary ore enrichment levels), whereas, apart from bacterial vestiges, fossil content is commonly low, and of low diversity with *Elphidium* and ostracodes as the most common representants. Lateral substitution of gypsum by limestone beds, as especially developed in the Carboneras area, is primarily the combined effect of common early microbial, and later karstification processes (Ch. 3.2).

The characteristic facies components and reconstruction of the basin geometry attest to deposition in a basin with predominant hypersalinity and common only moderately shallow depositional depths, where oxygen deficiency also was a common feature ('Environmental stage III') (Fig. 5.3). A cyclic sedimentation pattern ensued from periodic variations in these environmental parameters (Chs. 3, 4.2). Strongly hypersaline conditions during gypsum deposition alternated with moderately hypersaline values during sedimentation of laminated marls in the basin centres and stromatolites and oolitic sediments at its margins. At times only weakly hypersaline, nearly normal saline conditions are recorded in levels with typical marine fauna intercalated in both the basinal and marginal sequences (Ch 3; Riding et al. 1991a & b, Martin et al. 1993). Periodically, such conditions may even have been weakly brackish, according to the common weakly euryhaline character of the fauna and the presence of *Limnocythere* in a sample from the marginal facies (Ch. 3.3, unpublished data). Maximum waterdepths in the order of a 100-150 meters were still relatively common. Variations in this parameter are especially apparent in the sedimentation pattern in the marginal oolite sequence but the absence of unequivocal features attesting to subaerial or extremely shallow water conditions in the basinal sections indicates that such variations did not yet reach extreme values (Chs. 3, 4.2, Dronkert 1978, Beets & Roep 1978, Kleverlaan 1989). At least some episodes of warm-subtropical surface-water temperatures are indicated by a few intercalations of *Porites* reefs in the marginal carbonate facies.

The onset of Facies III deposition records a change from predominantly open, to predominantly 'restricted lagoon' conditions and is primarily marked by an increase in salinity (*'Event 3'* of Figs 5.1-,5.3). Although the first increase in salinity is relatively sharp, the onset of Stage III is still relatively gradual as seen from the appearance of intercalations with some of its characteristic components and parameter characteristics in the preceeding interval (Fig. 5.2). It is associated with a marginal hiatus near the base of 'Sequence B' as seen in common displaced reef blocks and a few terrigenous sand layers intercalated in marly sediments just below the evaporite series (Fig. 5.7; Chs. 3.2, 3.4).

A lower subinterval, associated with more persistent restriction and shallower depositional depths, and an upper one, attesting to relatively more open marine conditions, common deeper deposition and marginal onlap, can be recognized (Substages 'IIIa' & 'IIIb', respectively) (Figs 5.3 & 5.7, Ott D'Estevou 1980, Saint Martin & Rouchy 1990). The facies change which marks the onset of the latter conditions, at the base of the 'Upper Yesares Formation', is referred to as*'Event 4'* in Fig. 5.2 a.o..


EASTERN ALMERIA PROVINCE (SE SPAIN)
CENTRAL MEDITERRANEAN
MUELLER & HSU 1987
MUELLER 1993
BENSON EA. 1991
THIS STUDY
(CHAPTERS 3 & 4)
(STANDARD UNITS)
VERA BASIN
SORBAS BASIN
SORBAS-CARBONERAS AREA
CALTANISETTA BASIN/
DSDP 13 & 42A, ODP 107
CUE-VAS FM
CYPRIDEIS MARL
NOT STUDIED
CANOS FM
TURRE FM
ZORRERAS MEMBER (S.L.)
SORBAS MB
YESARES MB
YESARES MB (base)
ABAD MB
AZAGADOR MB
CHOZAS FM
ZORRERAS MB (s.s.)
YESARES MB
CUEVAS FM
Cypr mrl
FEOS FM
YESARES FM
UPPER
LOWER
UPPER SORBAS MB
LOWER SORBAS MB
U. MANCO LST.
GYP-SUM MB.
L. MANCO LST.
ABAD MRL
SAN TIA-GO MB
GYPSUM BRECCIA
AZAGADOR SST
base
TRUBI FM
ARENAZZOLO MB
PASQUASIA GYPSUM
(MAIN) SALT MB
C/D
A/B
CAL-CARE DI BASE
CATTOLICA GYPSUM
TRIPOLI MEMBER
MARNE A GLOBIGERINE
GESSOSO-SOLFIFERA FM
LICATA FM
UPPER EVAPORITES
LOWER EVAPORITES
PLIOCENE
MESSINIAN
late
middle
early
TORTONIAN
Ia
Ib
Ic
II

The paucity in fossils, the composition of their assemblages and the nature and abundance of the authigenic mineral assemblage attest to common deposition in a severely restricted basinal setting, but its largely predominant marine character is indicated by the intercalated marine fossiliferous levels and some stable isotope data on evaporites and carbonates (Longinelli 1980, Müller 1986). The characteristic components indicate predominant deposition in a semi-restricted to restricted, relatively deep, 'hypersaline lagoon' environment (Fig. 5.2). Water exchange with the open marine environment was difficult but at least inflow was basically continuous (Chs. 3.2, 3.3 & 4.2). Facies III has a number of features in common with modern hypersaline lagoons (Fig. 1.2) but differs by the common presence of subfacies attesting to relatively deep deposition. No modern examples of such deep hypersaline lagoons exist (Schmalz 1969, Dronkert 1985) but conditions in the Red Sea basin during the most intense Pleistocene 'glacial' episodes (Friedman 1972, McKenzie et al. 1980, Reiss & Hottinger 1981, Almogi-Labin et al. 1993) probably began to approach those of 'Stage III'.

Our Yesares Formation (Facies III) has many features in common with Central Mediterranean Lower Evaporites, especially, with middle Messinian evaporite-carbonate complexes of other shallower parts of this basin (e.g. Selli 1973, Sturani 1973, 1978, Delfaud & Revert 1974, Figliossi et al. 1978, Vai & Ricci Lucchi 1978, Orszag Sperber et al. 1980, Müller 1986, Saint-Martin & Rouchy 1990) and probably approximately correlates in age with these units (Chs 3.2, 3.3; Fig. 5.4). The deeper parts of the basin frequently received gravitatively displaced sulphate material (Schlager & Bolz 1977, Schreiber & Decima 1978, Vai & Ricci Lucchi 1978), commonly

<----- FIGURE 5.4. Proposed correlation of basinal lithostratigraphic units and major facies changes ('events') of the EAP to the Central Mediterranean Messinian standard stratigraphy (right) and comparison with earlier propositions (left). Numbers along columns refer to 'events' after Müller & Hsü 1987 for SE Spain (left) and Sicily (right). Our events (column 1c) refer to the major facies changes at the base of their 'facies events'. 6 and 7A have been merged into 6. Hatched intervals represent proposed major hiatuses.

In our concept (1C), major tectonic movements in the Entrance Area in the late Tortonian and subsequent rise of basin margins and infill of their centres in the early Messinian, lead to widespread restriction and, primarily, facies differentiation between SE Spain and the deeper parts of the Mediterranean (susbsequently suboxic versus anoxic bottom facies, massive gypsum deposition versus laminites, gypsum turbidites and massive rock salt, and shallow lacustrine facies with periodic carbonate and fluviatile intercalations vesrsus deeper lacustrine conditions with periodic sulphate evaporites, respectively). This approximately contemporaneous development of major environment types (restricted marine, evaporitic and continental) is best recorded in central parts of the inland Sorbas and Northern Nijar basins, where Messinian sedimentation is basically continuous, whereas major hiatuses, in which Mediterranean 'dessication' played an important role, are only recorded in coastal areas as the Vera Basin. Tectonics (primarily) and high global sealevel lead to approximately contemporaneous Pliocene inundation of the deeper parts of the Mediterranean and our deepest basinal areas and, subsequently to reinstallation of fully open marine conditions (events 8a and b).

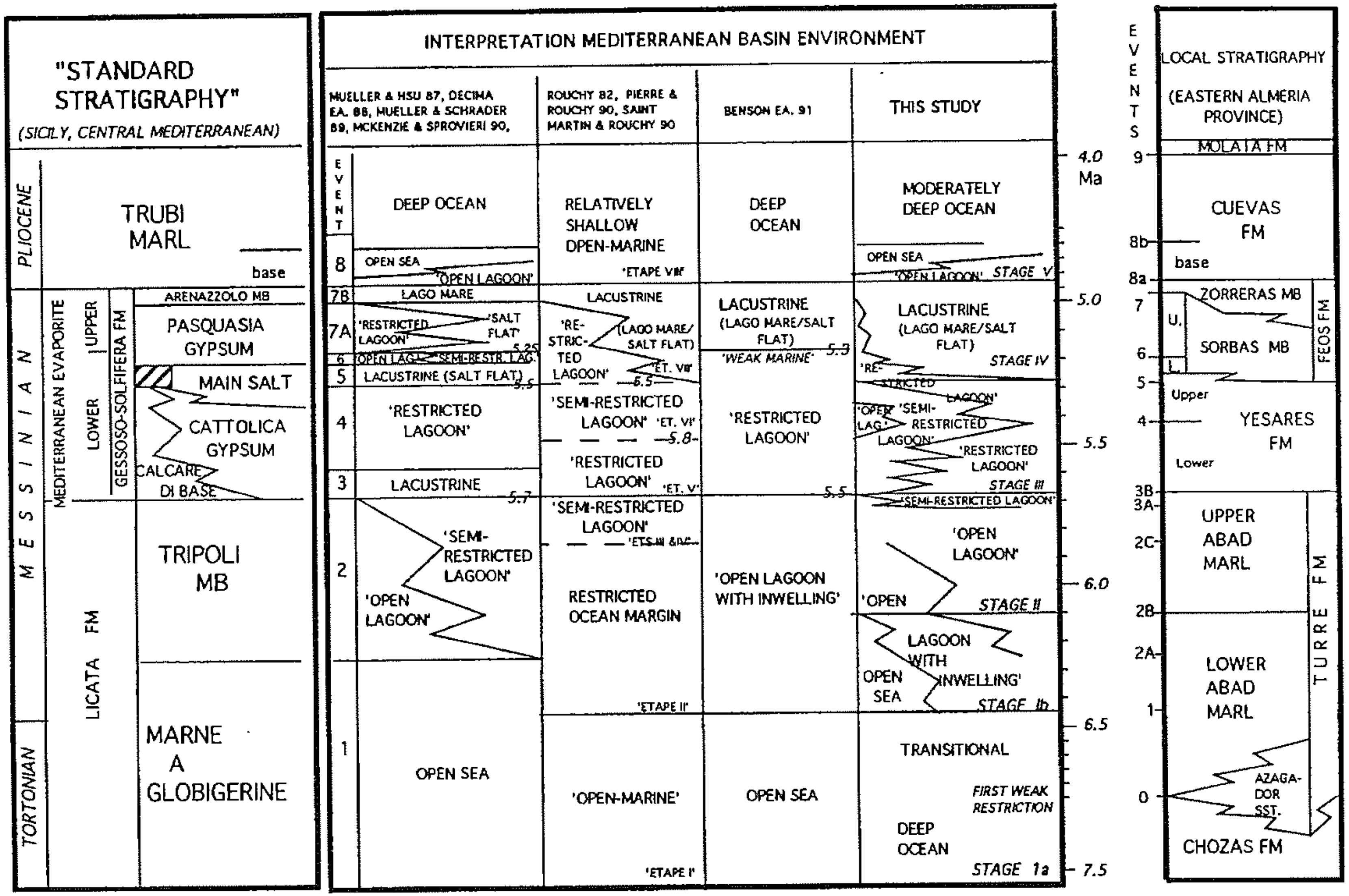
followed by an episode of major salt deposition of relatively short duration (Busson 1990; Fig. 5.4). A shelf of a, deep, hypersaline 'reflux' basin, with common but only relatively minor oscillations of the water level, is proposed as the predominant middle Messinian regional setting of the EAP (Figs 5.5 & 5.6). Some outflow of the shallower Mediterranean bottom waters probably was relatively common, and a few periods of 'open lagoonal conditions', both in relation with eustatic rise and possibly combined with increase in regional humidity and occasional estuarine circulation (cf. Müller & Schrader 1989; Fig. 5.6), still occurred in the middle-upper part of this interval.

The late Messinian is essentially represented by the 'Lower' and 'Upper Feos formation' of the Northern Nijar Basin, equivalent units in the Sorbas and Vera basins and Agua Amarga Breccia (pars) and sediments equivalent to the Upper Feos formation in the Agua Amarga Basin (Fig. 5.1). This interval is again best represented in the Northern Nijar and Sorbas basins, and only thin equivalents are present in the coastal areas. In the latter, features characterizing karstification and/or erosion of underlying Messinian deposits are the most obvious (Chs. 4.4 & 4.5, Cita et al. 1980, Clauzon 1980, Esteban & Giner 1980, Van de Poel 1980 and in Pineda Velasco et al. 1983). The late Messinian age of the the Feos deposits, is indicated by approximate stratigraphic continuity with earliest Pliocene deposits in the deepest basinal areas and detailed compatibility in facies development with the Late Messinian of North Italian sections (cf. Carbonnell 1978, Casati et al. 1978, Colalongo et al. 1978, Figliossi et al. 1978, Dondi & Rizzini 1978, 1980). Locally, however, in particular in the Sorbas and Agua Amarga basins, the upper part of the 'Zorreras Member' of the Feos Formation may be of Early Pliocene age (Ott D'Estevou 1980, Van de Poel et al. 1984; Fig. 5.1).

This interval is characterized by 'Facies IV' (Fig. 5.2): white, commonly laminated marl with numerous terrigenous clastic intercalations and the ostracode *Cyprideis* as the most common fossil. The autochthonous fossil content is somewhat higher than in the underlying interval but still relatively low, and of low diversity. Facies IV has a moderately high authigenic mineral content (calcite, dolomite, amorphous silica, mainly dispersed manganese and iron hydroxides and gypsum, halite and lignite). Cyclic facies development is still common.

The characteristic facies components attest to predominant deposition in moderately hypersaline to brackish basins where oxygen

FIGURE 5.5. Proposed scheme for Mediterranean Basin development with data from the EAP considered, as compared with earlier propositions. Environmental terminology as defined in Chapter 2. Further discussion in text. --------------->


"STANDARD STRATIGRAPHY"
(SICILY, CENTRAL MEDITERRANEAN)
PLIOCENE
MESSINIAN
TORTONIAN
TRUBI MARL
base
MEDITERRANEAN EVAPORITE
UPPER
LOWER
GESSOSO-SOLFIFERA FM
ARENAZZOLO MB
PASQUASIA GYPSUM
MAIN SALT
CATTOLICA GYPSUM
CALCARE DI BASE
TRIPOLI MB
LICATA FM
MARNE A GLOBIGERINE
INTERPRETATION MEDITERRANEAN BASIN ENVIRONMENT
MUELLER & HSU 87, DECIMA EA. 88, MUELLER & SCHRADER 89, MCKENZIE & SPROVIERI 90,
EVENT
DEEP OCEAN
OPEN SEA
'OPEN LAGOON'
LAGO MARE
'RESTRICTED LAGOON'
'SALT FLAT'
OPEN LAG.
SEMI-RESTR. LAG.
LACUSTRINE (SALT FLAT)
'RESTRICTED LAGOON'
LACUSTRINE
'SEMI-RESTRICTED LAGOON'
'OPEN LAGOON'
OPEN SEA
ROUCHY 82, PIERRE & ROUCHY 90, SAINT MARTIN & ROUCHY 90
RELATIVELY SHALLOW OPEN-MARINE
'ETAPE VIII'
LACUSTRINE
'RE-STRIC-TED LAGOON'
(LAGO MARE/ SALT FLAT)
'ET. VII'
'SEMI-RESTRICTED LAGOON'
'ET. VI'
'RESTRICTED LAGOON'
'ET. V'
'SEMI-RESTRICTED LAGOON'
'ETS III & IV'
RESTRICTED OCEAN MARGIN
'ETAPE II'
'OPEN-MARINE'
'ETAPE I'
BENSON EA. 91
DEEP OCEAN
LACUSTRINE (LAGO MARE/SALT FLAT)
'WEAK MARINE'
'RESTRICTED LAGOON'
'OPEN LAGOON WITH INWELLING'
OPEN SEA
THIS STUDY
MODERATELY DEEP OCEAN
OPEN SEA
'OPEN LAGOON'
STAGE V
LACUSTRINE (LAGO MARE/SALT FLAT)
STAGE IV
'RE-STRICTED
LAGOON'
'OPEN LAG.'
'SEMI-RESTRICTED LAGOON'
'RESTRICTED LAGOON'
STAGE III
'SEMI-RESTRICTED LAGOON'
'OPEN LAGOON'
'OPEN
STAGE II
LAGOON WITH INWELLING'
OPEN SEA
STAGE Ib
TRANSITIONAL
FIRST WEAK RESTRICTION
DEEP OCEAN
STAGE 1a
4.0 Ma
5.0
5.5
6.0
6.5
7.5
EVENTS
LOCAL STRATIGRAPHY
(EASTERN ALMERIA PROVINCE)
MOLATA FM
CUEVAS FM
base
ZORRERAS MB
SORBAS MB
FEOS FM
Upper
YESARES FM
Lower
UPPER ABAD MARL
LOWER ABAD MARL
TURRE FM
AZAGA-DOR SST.
CHOZAS FM

deficiency still played an important role (Fig. 5.3). Intercalations of fluviatile deposits represent occasional fresh-water conditions. Periodic well-oxygenated conditions become apparent from limonitic and burrowed marls associated with the clastic intercalations. Thin-bedded limestone intercalations in the middle part of this interval (Upper Sorbas Member) in the Northern Nijar Basin compare in facies and stratigraphic setting with the Late Messinian so-called 'Colombacci' limestones of NE Italy and probably likewise represent deposition in alkaline waters (Colalongo et al. 1978). Paucity in fossils, sedimentological data and analyses of the basin geometries indicate common depositional depths of several tens of meters in particular in the earlier part of 'Stage IV' (Chs. 3.2 & 3.3, Pagnier 1978, Roep et al. 1979).

Rapid and often extremely strong fluctuations of environmental parameters are common (Fig. 5.3). Severely hypersaline conditions are recorded in a few levels with evaporite minerals. Moderately hypersaline to moderately brackish waters are probably represented by levels with only fish remains and/or *Cyprideis* and the foraminifer *Ammonia tepida* (Ch. 3.3, Anadon 1992, Boomer 1993). Hyper-brackish to oligohaline conditions are represented by levels with common remains of *Chara, Dreissena, Limnocardium* and a more diverse ostracode fauna (Chs. 3.3, 4.4, Ott D'Estevou 1980). Periodic, extremely shallow and even subaerial conditions are recorded in littoral, shallow lagoonal, fluviatile and calcretic intercalations in or near the most central parts of the inland Sorbas and Northern Nijar basins (Chs. 3.2, 4.2), whereas at least one or two episodes of major calcrete formation, karstification and/or river incision and localized valley fill are recorded in the coastal basins (Chs. 4.4-4.5; Figs 5.1, 5.8, 5.9).

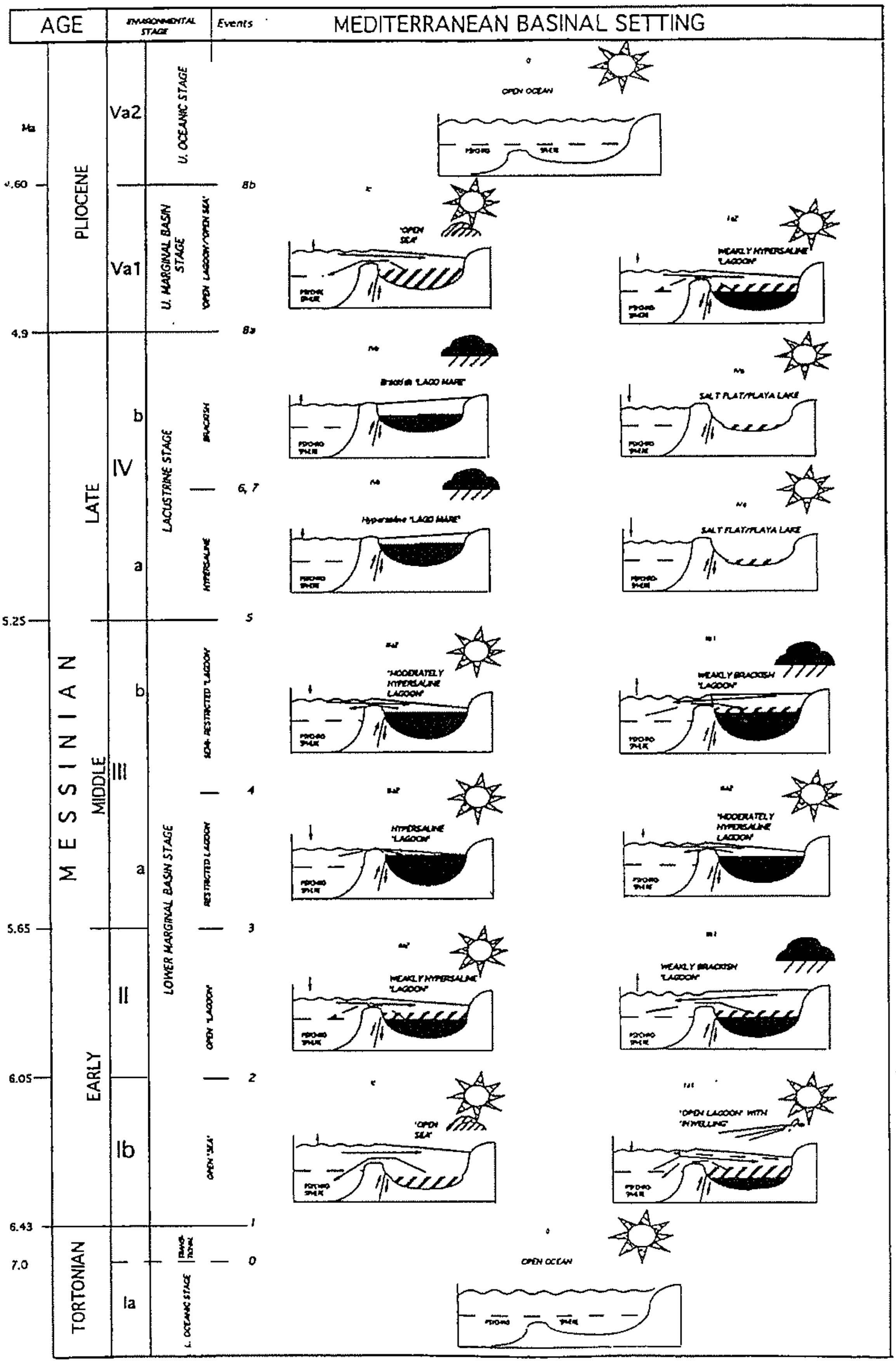
The onset of 'Stage IV' is marked by the appearance of clastic intercalations and a decrease in authigenic evaporitic mineral and marine fossil content. This facies change is referred as *'Event 5'* in Fig. 5.2 a.o.. Intercalations with reworked evaporites in the Sorbas and Northern Nijar basins (Pagnier 1978, Ch. 3.2) indicate periodic subaerial erosion of outer parts of their centres at this moment. A major erosion event associated with and preceeding deposition of the 'Gypsum breccia' in the Vera Basin probably also dates from this episode (Montenat et al. 1976, Van Hoeflaken 1984, Dronkert 1985, Sennema 1986, Barragan et al. 1990, partially reinterpreted: Fig. 5.1). The tendency towards lower salinities, shallower water and better oxygenation basically pursues through Stage IV and Events *'6'* and *'7'* represent further distinct steps in, what is otherwise recorded as, a relatively gradual development in the central parts of the inland Sorbas and Northern Nijar basins (Figs. 5.2 & 5.3).

An increase in fossil content with the appearance of typically brackish 'lago mare' components characterizes Event 6 at the base of the 'Upper Feos Formation', the facies of which represents 'Substage IVb' (Figs 5.2 & 5.3). Event 6 is associated with common smaller hiatuses at the margins, and near the central part of the Sorbas Basin (Ott D'Estevou

1980, Dronkert 1985, pers. comm. Th.B. Roep 1993), whereas unconformities probably also developed at the base of this substage near the margins of the Northern Nijar Basin (Ch. 3.2; Figs 5.7-5.9). The hiatus at the base of the '*Cyprideis* marl' of the Vera Basin is at least partially due to erosion related with Event 6, which may be responsable for the irregular distribution of the Gypsum breccia, leaving only erosional remnants of a, formerly larger, 'valley fill' (Fig. 5.1; Ch. 4.4).

A last, important Messinian facies change (Event 7), is recorded near the top of this interval, where it represents the onset of predominant hyperbrackish lacustrine, fluviatile and subaerial conditions represented by the typical 'Zorreras' deposits (Subfacies IVc) . This event is most obvious in the Sorbas Basin and may occur later in the central part of the Northern Nijar Basin, where it can barely be recognized in its eastern part (Fig. 5.1). The appearance of hyperbrackish to oligohaline ostracodes in the top of the *Cyprideis* marl of the Cuevas del Almanzora section (Ch. 4.4, Cita et al. 1980), probably represents this event in the Vera Basin. The onset of typical Zorreras deposition is again associated with erosional unconformities in the outer parts of the centre of Sorbas and Northern Nijar basins (Figs 5.7 & 5.8; Offringa 1991, Ch. 3.2). In the northern-central part of the Agua Amarga Basin, a local incised valley fill of reddish sands with *Microcodium* and conglomerates cuts down till a karstified selenitic gypsum bed from the base of the Yesares Formation (Van de Poel et al. 1984, Dronkert 1985; Fig. 5.1). This strongly suggests that at least part of the river incision and karstification of evaporites and carbonates in this area was either associated with, or slightly preceeded Event 7. The extremely poor development of Zorreras deposits in the Agua Amarga Basin on the other hand indicates still a subsequent erosion event in this area. This may be completely ascribed to erosion related with tectonic movements at the base of the middle Pliocene Molata Formation ('Event 9', see hereafter), but the existence of a new erosion event at the very top of the Messinian cannot be excluded. Stage IV thus was characterized by repeated erosional events related to periodic drawdown of the 'sea' level, but major erosion probably concentrated around 'Event 6'.

Relatively common lamination, low fossil diversity and extremely variable abundancy, as well as the authigenic mineral content, attest to a restricted environment in the late Messinian also (see also Benson 1978a). According to the fossil components, this may have either been an extremely restricted lagoon or a restricted lake (cf. Caspers 1957, Zenkevitch 1957, Archambault-Guezou 1976, Perthuisot & Guelorget 1987, Anadon 1992, Boomer 1993), but essentially lacustrine conditions are indicated by the virtual or complete absence of (autochthonous) marine fossils (Fig. 5.2). Environmental stage IV can be considered to represent subsequently predominant hypersaline and brackish lacustrine conditions with a distinct tendency to fluviatile-deltaic deposition in its later part (Figs 5.1-5.3).



AGE
ENVIRONMENTAL STAGE
Events
MEDITERRANEAN BASINAL SETTING
Ma
4.60
4.9
5.25
5.65
6.05
6.43
7.0
PLIOCENE
MESSINIAN
LATE
MIDDLE
EARLY
TORTONIAN
Va2
Va1
IV
b
a
III
b
a
II
Ib
Ia
U. OCEANIC STAGE
U. MARGINAL BASIN STAGE
LACUSTRINE STAGE
LOWER MARGINAL BASIN STAGE
L. OCEANIC STAGE
8b
8a
6, 7
5
4
3
2
1
0
OPEN OCEAN
'OPEN SEA'
WEAKLY HYPERSALINE 'LAGOON'
Brackish 'LAGO MARE'
SALT FLAT/PLAYA LAKE
Hypersaline 'LAGO MARE'
'MODERATELY HYPERSALINE LAGOON'
WEAKLY BRACKISH 'LAGOON'
HYPERSALINE 'LAGOON'
'OPEN LAGOON' WITH 'INWELLING'

The characteristic succession in the EAP of a lower interval with poorly fossiliferous strata with common redeposited evaporites andbreccias, locally represented by a more or less important hiatus, followed by an interval which contains the typical lago mare fauna and often associated with more or less important fluviatile-deltaic intercalations towards its top, compares with that of numerous sections from more internal parts of the Mediterranean (e.g. Decima & Sprovieri 1973, Decima & Wezel 1973, Sturani 1973, 1978, Fortuin 1977, Cita et al. 1978, Colalongo et al. 1978, Dondi & Rizzini 1978, Figliossi et al. 1978, Mascle & Heimann 1978, Vismara Schilling et al. 1978, Cita & Colombo 1979, Orszag Sperber et al. 1980, Van Hinte et al. 1980, Rouchy 1981, 1982; Fig. 5.4).). Our data confirm the presence of a lake of enormous dimensions during the late Messinian episode (Ruggieri 1967, Ruggieri & Sprovieri 1976, Cita et al. 1978, Benson et al. 1991; Figs. 5.5, 5.6). A relatively high salinity level was probably esentially maintained through the reworking of older evaporites (Ruggieri & Sprovieri 1976, Van de Poel 1991). Periodic climatic variations played an important role in the sedimentary facies development and variations in water-budget of the basin (Chamley & Robert 1980, Van Hinte 1990, 1991, Benson et al. 1991). This is confirmed by facies variations in our area, where calcrete horizons and common *Microcodium* represent periodic relatively warm and dry conditions (Roep & Beets 1977, Klappa 1978), whereas widespread lake and karst development attest to relative humidity. Characteristic facies components, in particular the benthic foraminifer, ostracode and mollusk assemblages, are highly comparable to those of modern Lake Aral (Ouzbekistan, ex USSR) (cf. Zenkevitch 1957, Anadon 1992, Boomer 1993). Most striking parallels exist between the late Messinian Mediterranean environment and that of Lake Aral in its most recent development of major shrinking and local evaporite deposition, which is here the result of man-induced changes in the local climate (Boomer 1993, Heeres 1994).

The hypotheses of one or more important drawdowns, and probable dessications, of the Mediterranean 'lago mare' during periods of strongly negative water budget (Van Hinte 1991, 1992, Ch. 2) are confirmed by the important erosional unconformities, with local river incision in the coastal Vera and Agua Amarga basins (Chs. 4.4 & 4.5, Cita et al. 1980, Clauzon 1980; Figs 5.6.IVa, 5.8 & 5.9). Rapid sedimentary infill during the Messinian, as also recorded in our area, probably was a common process

<----- FIGURE 5.6. Cartoon of the proposed development of the regional setting of EAP and Mediterranean(cf. Fig. 1.2). Predominant basinal settings for each (sub)episode to the left.

An evolution, which slowly evolved from 'oceanic' (stage Ia) in the Tortonian, via 'open sea' (stage 1b) in the earliest Messinian, 'open lagoon' (stage II) in the late early Messinian , and '(semi-)restricted lagoon' (stage III) in the middle Messinian, to lacustrine (stage IV) in the late Messinian, and a rapid return to open lagoon/open sea (stage Va1) and then to oceanic (stage Va2) in the Pliocene is proposed. Facies development further depends on the 'estuarine/anti-estuarine mode' of the basin (Ch. 2).

in the Mediterranean (Montadert et al. 1978, Rouchy 1982, fig. 9, Van der Zwaan 1982, Busson 1990; Figs. 5.7-5.9), but benthic foraminifer and geophysical data for early Pliocene marls from the its deepest parts (Sprovieri 1974, 1977, Brolsma 1978, Wright 1978b, Génesseaux & Lefebvre 1980) still suggest maximum depths in the order of one to two kilometers. During evaporitic drawdown, fresh-water supply was probably sustained for a considerable time by the emptying of the karstic reservoirs contained in the landmasses surrounding the basin, whereas even capturing of higher-lying lakes may have taken place (Ruggieri & Sprovieri 1976, Hsü et al. 1977, 1978, Ryan & Cita 1978, Carbonnell 1980). Relative persistence of submerged conditions through release of karst waters can presently be observed in seasonally dessicated lagoons of the NW Mediterrean margin (Van de Poel & Boekschoten, in prep.).

Conditions during the **Pliocene** in the deeper basinal areas of the EAP are represented by the Cuevas, Molata and Espiritu Santo formations. This interval is best represented in the Vera Basin, whereas it is poorly developed in the Basin of Sorbas (Fig. 5.1). Pliocene age is indicated by the presence of the Mediterranean *Globorotalia margaritae* and *G. puncticulata* (MPl 2 & 3) planktonic foraminifer zones in the Vera and Nijar basins, where probable also the older *Sphaeroidinellopsis* Zone (MPl1) and the younger MPl 4 are represented, and by Pliocene bivalves in the Sorbas Basin (Chs. 3 & 4.4; Ott D'Estevou 1980, Postma 1978, unpublished data; Figs. 5.1 & 5.3).

Bioturbated, calcareous silts, sandstones and conglomerates with common *pectinids* is the dominant lithology of 'Facies V', which is further characterized by abundant and diverse remains of other marine organisms and an authigenic mineral content essentially represented by dispersed glauconite grains (Fig. 5.2).

The facies components record predominant open-marine conditions with relatively normal salinities and oxygen-content of the bottom-waters during 'Environmental stage V' (Figs 5.2 & 5.3). Outer neritic to uppermost bathyal deposits are still formed during the oldest part of this interval, whereas further shoaling to inner neritic depths occurred in the course of the Pliocene (Fig. 5.3). Subtropical surface-water temperatures are still indicated by the abundance of representatives of the *Globigerinoides obliquus* and *G. trilobus* groups of planktonic foraminifers (cf. Ch. 4.3, Thunnell et al. 1991). The absence of other subtropical surface water elements (*Globorotalia menardii*, hermatypic corals and other associated carbonate platform biota) on the other hand first suggests a certain cooling in respect to the Late Miocene. Alternatively, this feature may only reflect cooling of Atlantic surface waters to the west of the Mediterranean Entrance Area, where such biota also disappeared at the end of the Miocene (Zachariasse & Spaak 1983, Benson et al. 1991, pers. comm. G. Boekschoten 1991) and hence the physical impossibility to reenter the Mediterranean Basin for these organisms. Thus, in the Mediterranean and nearby areas an 'extinction event' is recorded, whereas in other regions the Mio-Pliocene boundary interval is rather characterized by speciation of planktonic foraminifers and larger mammals as elephantids and hominids (Dowsett 1989, Cooke & Maglio 1972, Howell 1972, Pilbeam 1972).

The oldest part of this interval is best documented in the Vera Basin, while continental conditions characteristic of Stage IV possibly persisted in some of the higher basinal areas. At least an episode of early Pliocene continental conditions, associated with 'Event 9', is recorded in the Carboneras area, where it probably contributed to karstification and erosion of Messinian deposits(ch. 4.5; Figs 5.1, 5.3, 5.7-5.9). Still somewhat restricted marine, typical marginal basin conditions during deposition of the base of the early Pliocene Cuevas Formation are indicated by reduced marine microfossil diversity and fluctuations in its composition (Figs. 5.2, & 5.3; Ch. 3.3 and data from Montenat et al. 1976, Gonzalez Donoso & Serrano 1978, Cita et al. 1980 and Geerlings & Van de Poel unpublished). Similar features and even the absence of benthos have been recorded in the very base of the Pliocene in sections from the Central Mediterranean (Sprovieri 1978, Hasegawa et al. 1990; McKenzie & Sprovieri 1990) and further attest to a setting evolving from 'open lagoon' to 'open sea' with slightly aberrant salinities and some oxygen-deficiency and probably a relatively elevated nutrient content of the bottom waters and cyclic variations (Figs 5.2, 5.5 & 5.6).

The reappearance of high-diverse marine microfossil assemblages indicates the onset of fully open marine conditions somewhat above the base of the Pliocene (Figs 5.2 & 5.3). Some of its elements (common *Oridorsalis* and the keeled globorotaliid *G. margaritae*) suggest conditions different from those in the present Mediterranean (cf. Cita 1975c, Duprat 1983, Hasegawa et al. 1990). Contemporaneous facies from the Central Mediterranean, which contain typically 'psychrospheric' elements, further indicate an oceanic setting for the Mediterranean Basin during most of the Pliocene (Benson 1978, Van Harten 1983, Thunnell et al. 1987, Hasegawa et al. 1990, McKenzie & Sprovieri 1990; Fig. 5.6).

THE LOCAL GEODYNAMIC DEVELOPMENT

In all the basinal areas, maximum depositional depths during the Pliocene were distinctly less than during the earliest Messinian which attests to the importance of local tecto-sedimentary processes in the intervening Mio-Pliocene boundary interval, as do local marginal angular unconformities and a general progradation tendency and a basinward shift in 'coastal onlap', between latest Tortonian-earliest Messinian 'Globigerina marls' of 'Sequence A' (Turre Fm) in respect to Pliocene 'Pectinid sands ('Sequence D'/Molata Fm) (Chs. 3 & 4.5, Figs. 5.3, 5.7-5.9). This appears primarily as the result of differential uplift of the basin-margins and sedimentary infill of their centres, whereas in particular in the Agua Amarga Basin, and, to a much lesser extent, also in the Basin of Vera, general uplift is recorded.

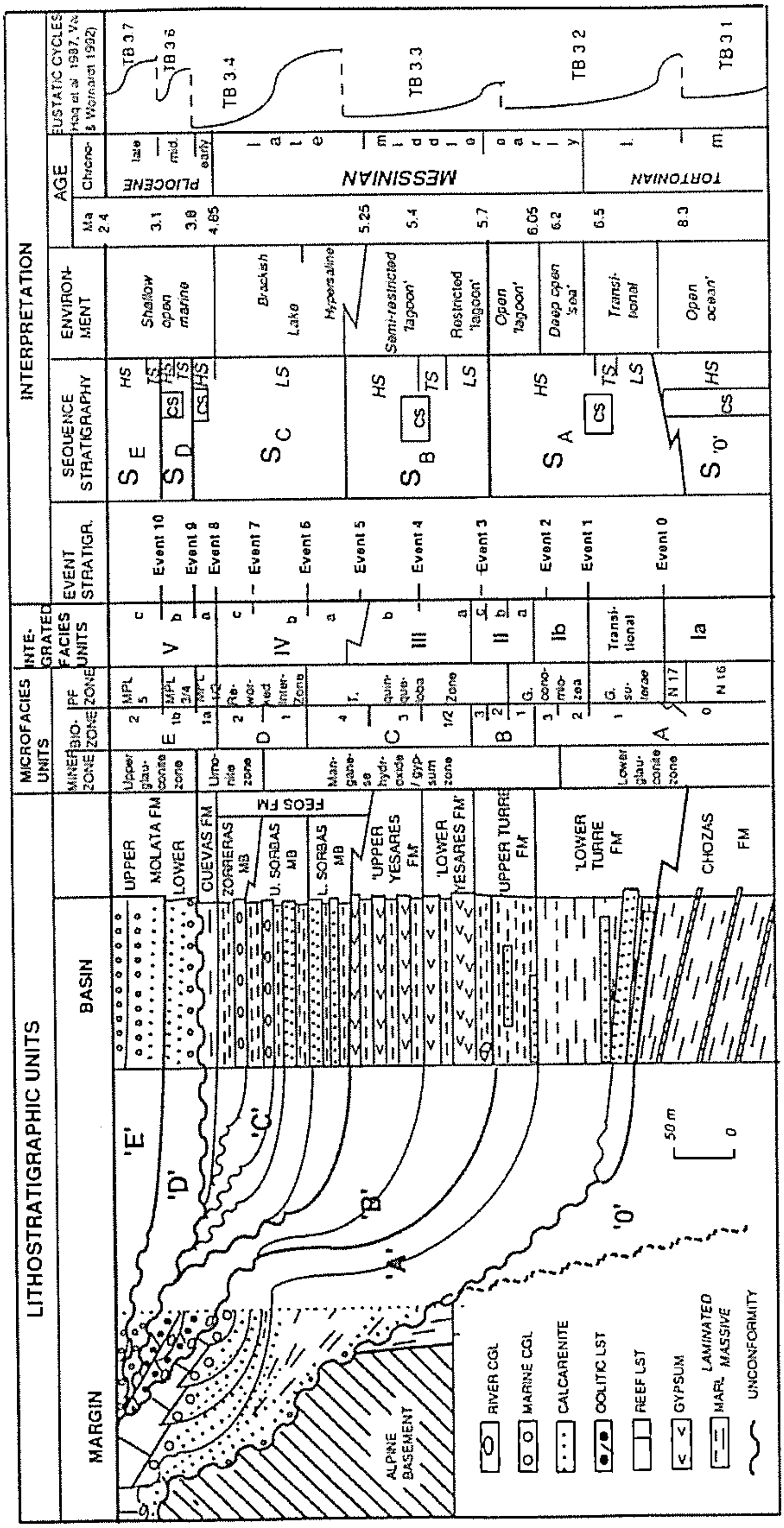
An important phase of relative shoaling, associated with major changes in local paleogeography is already recorded in the Late Tortonian with '*Event 0*'. During the earlier part of the Tortonian and older Neogene periods, deep, relatively elongated troughs existed, related to the E-W directed wrench fault systems of the area, whereas towards the end of the Tortonian, shallower more invidualized basins with depocenters related to more N-S directed faults came into existence (Völk 1967, Montenat et al. 1990, Boorsma 1993, Ch. 2). Benthic fossil assemblages indicate middle bathyal depositional depths of many hundreds of meters for the middle Tortonian Gatar Marls of the Sorbas, Tabernas, Vera and Nijar basinal areas of the Almeria Province, whereas upper bathyal depths of a few hundred meters are recorded in the latest Tortonian-earliest Messinian (Ott D'Estevou 1980, Troelstra et al. 1980, Kleverlaan 1989, Fig. 5.3).

The paleogeographic restructuration of the area becomes first apparent in the distribution of common marginal angular unconformities and of the associated lowstand and transgressive deposits of the Azagador Sandstone at the base of the Turre Formation and is further reflected in the distribution of its basinal Abad Marls. In the northern part of the area transgression upon the Sierra de los Filabres massif is observed and the more localized depocenters of the Vera and Sorbas basins became situated closer to this structure (Volk 1967, Coppier et al. 1990, Fig. 2.3). To the south, the western end of the Sierra Alhamilla

FIGURE 5.7. Interpretation of the Mio-Pliocene of the central part of the Nijar Basin and its margin in terms of sequence stratigraphy. A number of depositional sequences can be recognized on basis of more or less well-developed condensed sections (CS), separated by lowstand deposits, in the basin centre, and correlated with onlapping sediments and unconformities in its shallower parts (cf. Vail & Wornardt 1992).

The major unconformities at the base of 'S_0', S_A and S_D are primarily of tectonic origin, whereas that at the base of S_C is primarily due to Mediterranean 'sealevel' lowering as the result of a combination of a tectonically elevated sill system, global lowstand and periodic dry Mediterranean climate conditions.

The (third order) sequences make out part of a higher order cycle which is reflected in a, somewhat fluctuating but consitent, tendency to progradation attested by coarsening-upward and reduction in waterdepth indicated by the fossil assemblages. The Messinian regression event (progradation and restriction, which is also reflected in its authigenic mineral content), separates sediments representing basinal portions of systems tracts (below) from those representing shelf portions of systems tracts (above). ------------>


LITHOSTRATIGRAPHIC UNITS
MARGIN
BASIN
MICROFACIES UNITS
MINER. ZONE
BIO-ZONE
IPF ZONE
INTEGRATED FACIES UNITS
INTERPRETATION
EVENT STRATIGR.
SEQUENCE STRATIGRAPHY
ENVIRONMENT
AGE
Ma
Chrono-
EUSTATIC CYCLES
UPPER MOLATA FM
LOWER
CUEVAS FM
ZORRERAS MB
U. SORBAS MB
L. SORBAS MB
FEOS FM
'UPPER YESARES FM'
'LOWER YESARES FM'
UPPER TURRE FM
'LOWER TURRE FM'
CHOZAS FM
ALPINE BASEMENT
RIVER CGL
MARINE CGL
CALCARENITE
OOLITIC LST
REEF LST
GYPSUM
LAMINATED
MARL MASSIVE
UNCONFORMITY
50 m
0
Event 10
Event 9
Event 8
Event 7
Event 6
Event 5
Event 4
Event 3
Event 2
Event 1
Event 0
Shallow open marine
Brackish
Lake
Hypersaline
Semi-restricted 'lagoon'
Restricted 'lagoon'
Open 'lagoon'
Deep open 'sea'
Transitional
Open ocean'
PLIOCENE
MESSINIAN
TORTONIAN
TB 3.7
TB 3.6
TB 3.4
TB 3.3
TB 3.2
TB 3.1

became a prominent positive structure, as well as the SW part of the Sierra de Gata, whereas the Agua Amarga Basin came into existence by transgression upon its NE part (Fig. 2.2 for localities). This is probably related to a change in the direction of the compressional field exerted on the area (Montenat et al. 1990). Marked, permanent decrease in subsidence, in gravitative sedimentation and in volcanic activity with Event 0, strongly suggest that it also marks a decrease in tectonic activity.

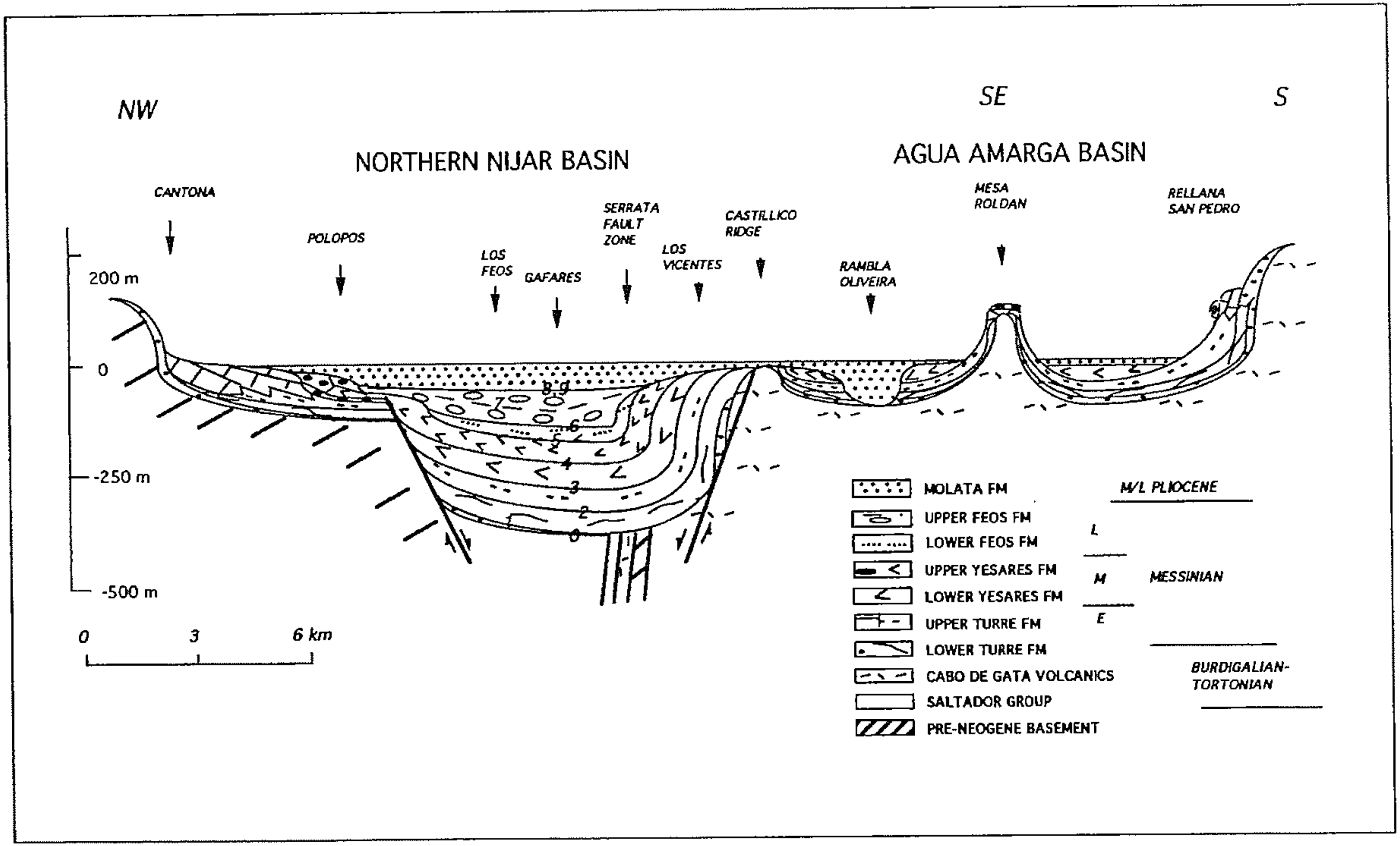
The relatively important transgression in the top of the Tortonian must still primarily be related with the tectonic movements (Ch. 3.4; Busson 1990). During this transgression all of the basins of the EAP, aswell as their interconnecting corridors became covered by a moderately shallow sea, which still largely extended their present margins (e.g. Figs. 2.3 & 5.8).

Shoaling and progradation of basin margins between the earliest Messinian and the Pliocene is partially the effect of sedimentary infill, as particularly seen in the centre of the Sorbas and Northern Nijar basins, where a thick intercalated shoaling-upward serie is present (Figs 5.3, 5.7-5.9). Common tectonic reduction of 'accomodation space' in the Mio-Pliocene boundary interval however also played an important role in the facies and paleogeographic changes, as becomes especially apparent from angular unconformities and abrupt vertical facies changes between directly superimposed deeper, earliest Messinian and shallower marine Pliocene deposits at the margins of the Nijar and Tabernas basins, as well as more or less similar, abrupt facies changes in the more central parts of the Agua Amarga and Vera Basins (Chs. 3 & 4.5; Völk 1967, Montenat et al. 1976, 1990, Kleverlaan 1989, Boorsma 1993; Figs 5.8 & 5.9).

Part of the 'Mio-Pliocene tectonics' took place within the early Pliocene as indicated by small angular differences and sealing of fault contacts observed at the base of the middle Pliocene Molata sediments ('Sequence D') of the Carboneras area (Ch. 5.5; **Event 9** of Fig. 5.8 a.o.) and of comparable sediments in the Almeria arae (Iaccarino et al. 1975). From the southern Nijar Basin, Boorsma (1993) reported a rotational unconformity between units, which probably correlate with our Cuevas

FIGURE 5.8. Reconstructed basin configuration of the Carboneras area towards the end of the Pliocene (simplified, in particular for the Serrata Fault Zone area). Vertical exaggeration: 12 x. See Figs. 2.3, 3.2.1 and 4.5.1 for locations. Numbers refer to main facies changes, the 'events'.

Note the difference betweenthe relatively subsident, inland Northern Nijar Basin, where Messinian sedimentation was more or less continuous and sedimentary facies well-developed and preserved in its central part, and the coastal Agua Amarga Basin, where the late Messinian is poorly developed and deposits from its middle part strongly altered, and where a wide 'proto-Oliveira valley' was incised due to a combination of major late Messinian 'sealevel' drawdown and uplift in the Mio-Pliocene boundary interval. The valley later served as a 'Pliocene ria' as described by Clauzon (1980 a.o.) from other margins of the Western Mediterranean. ------------------->


NW
SE
S
NORTHERN NIJAR BASIN
AGUA AMARGA BASIN
CANTONA
POLOPOS
LOS FEOS
GAFARES
SERRATA FAULT ZONE
LOS VICENTES
CASTILLICO RIDGE
RAMBLA OLIVEIRA
MESA ROLDAN
RELLANA SAN PEDRO
200 m
0
-250 m
-500 m
0
3
6 km
MOLATA FM
UPPER FEOS FM
LOWER FEOS FM
UPPER YESARES FM
LOWER YESARES FM
UPPER TURRE FM
LOWER TURRE FM
CABO DE GATA VOLCANICS
SALTADOR GROUP
PRE-NEOGENE BASEMENT
M/L PLIOCENE
L
M
E
MESSINIAN
BURDIGALIAN-TORTONIAN

and (Lower) Molata formations, whereas Völk (1967) proposed tectonic movements in between the deposition of comparable 'Lower' and 'Upper Cuevas' sediments in the Vera basin.

The major basinward shift in coastal onlap and reduction in depositional depth without sedimentary infill in the basin centre, which took place between the early-middle Messinian and deposition of the marine Pliocene in the Agua Amarga Basin, are not observed near the more 'passive' margins of the EAP (north to NW margins of the Vera and Nijar and N and SW margins of the Sorbas Basin) (ch. 5.5; Figs 2.3, 5.7-5.9). This suggests that these tectonic movements were differential and concentrated near the coastal wrench-fault zones. Although considered as esssentially early Pliocene of age, the existence of a late Messinian component cannot be excluded on basis of the present data.

Facies differences and marginal angular unconformities between the highstand deposits of the early Messinian 'Turre Sequence' ('A') and those of the early Pliocene Cuevas Formation ('Sequence C') in the Vera Basin, which are relatively smaller but similar in outbuilding to those recorded at the base of the Molata deposits of the Carboneras area, strongly suggest that part of the 'Mio-Pliocene tectonics' took place during the Messinian. Again, here there may be a late Messinian component, but a number of data suggest that an important part of this unconformity is of earlier Messinian age.

Tectonic movements related with early Messinian Event 3, were earlier proposed by Völk (1967), Montenat et al. (1976), and Kleverlaan (1989) for different parts of the Almeria Province, whereas at least additional importance of sedimentary infill during the biologically productive 'Stage II' was suggested in Troelstra et al. (1980). These processes can explain a number of stratigraphic and facies features of the Mio-Pliocene interval. In the western Vera Basin, where the vertical facies shift between the latest early Messinian Cantera Reef deposits and marginal sediments of the Cuevas Formation is relatively minor, a major facies shifts exist between the earliest Messinian and the early Pliocene, whereas 'Veritic' volcanic eruptions of this area probably are of latest early Messinian age (Völk 1966a, 1967, unpublished data of H. Dronkert, L.P.A. Geerlings and the author).

Tectonic movements in this interval probably also were a factor in the important, but extremely gradual, out and downbuilding of the marginal Cantera Limestone, described by Völk 1966a, Esteban et al. 1977, Dronkert 1977b, Pagnier 1977, Ott D'Estevou 1980, Esteban & Giner 1980, Dabrio et al. 1981, Van de Poel et al. 1984, etc. from our area and by e.g. Gourinard (1955) and Saint-Martin & Rouchy (1990) from other parts of the Western Mediterranean. This feature is presently commonly primarily explained as the result of exaggerated Mediterranean drawdown, whether or not associated with eustatic lowering (Dabrio et al. 1981, Saint-Martin & Rouchy 1990, Busson 1990), but such drawdown is, in any case, difficult to envisage without previous tectonic reduction of sill-depth to a critical level. Tectonics are also suggested by the presence of regressive conditions in our area and the Mediterranean in an interval where global sealevel highstand is commonly proposed (Van Gorsel & Troelstra 1980, Haq et al. 1987, Vail & Wornardt 1991; Fig. 5.7), and the reduction in depositional depth, basinward shift in coastal onlap and strong reduction in fossil diversity between the condensed sections and associated marginal highstand deposits of 'Sequences A and B' (Figs. 5.3, 5.7). An angular unconformity within the early Messinian has been reported from the Fortuna Basin (Lukowski et al. 1987), whereas the abundance of Cretaceous/Early Tertiairy planktonic forams in the Santiago Turbidite Member of the Vera basin (Fig. 2.1) strongly suggests uplift of the "Subbetic hinterland" to the north (Völk 1967, Montenat et al. 1976, personal observations; Fig. 2.1).

As the result of the early Messinian sedimentary infill and tectonics, 'hypersaline lagoon' deposition during the middle Messinian already took place in somewhat shallower and smaller basins, which were however still largely interconnected as indicated by the facies development in or near a number of the 'corridors' and the importance of later upift of oblique 'ridges' in the latter (Ch. 4.5, Harvey & Wells 1987, Figs 2.3, 5.3, 5.7-5.9).

Tectonics near the base of the late Messinian probably were at least an additional factor in the production of the polymict olisthostrome (Gypsum breccia) in the Vera Basin (Montenat et al. 1976, Van Hoeflaken 1984, Barragan et al. 1990), whereas, although the Agua Amarga Breccia is primarily considered the result of evaporite karst, preceeding faulting probable facilitated this karst development (cf. Simpson 1988, Warren et al. 1990). The extremely reduced youngest Messinian deposits in the Agua Amarga Basin, as well as the absence of its deeper water facies ('Upper Sorbas Member') suggests that this area was already subjected to differential uplift. This probably enhanced river incision in the, more open, northern part of the basin, which, in its turn, will have further enhanced karst formation. As in the early Pliocene, late Mesinian movements may have been essentially concentrated near the eastern coastal fault zones of the area (see previous discussion of 'Event 9').

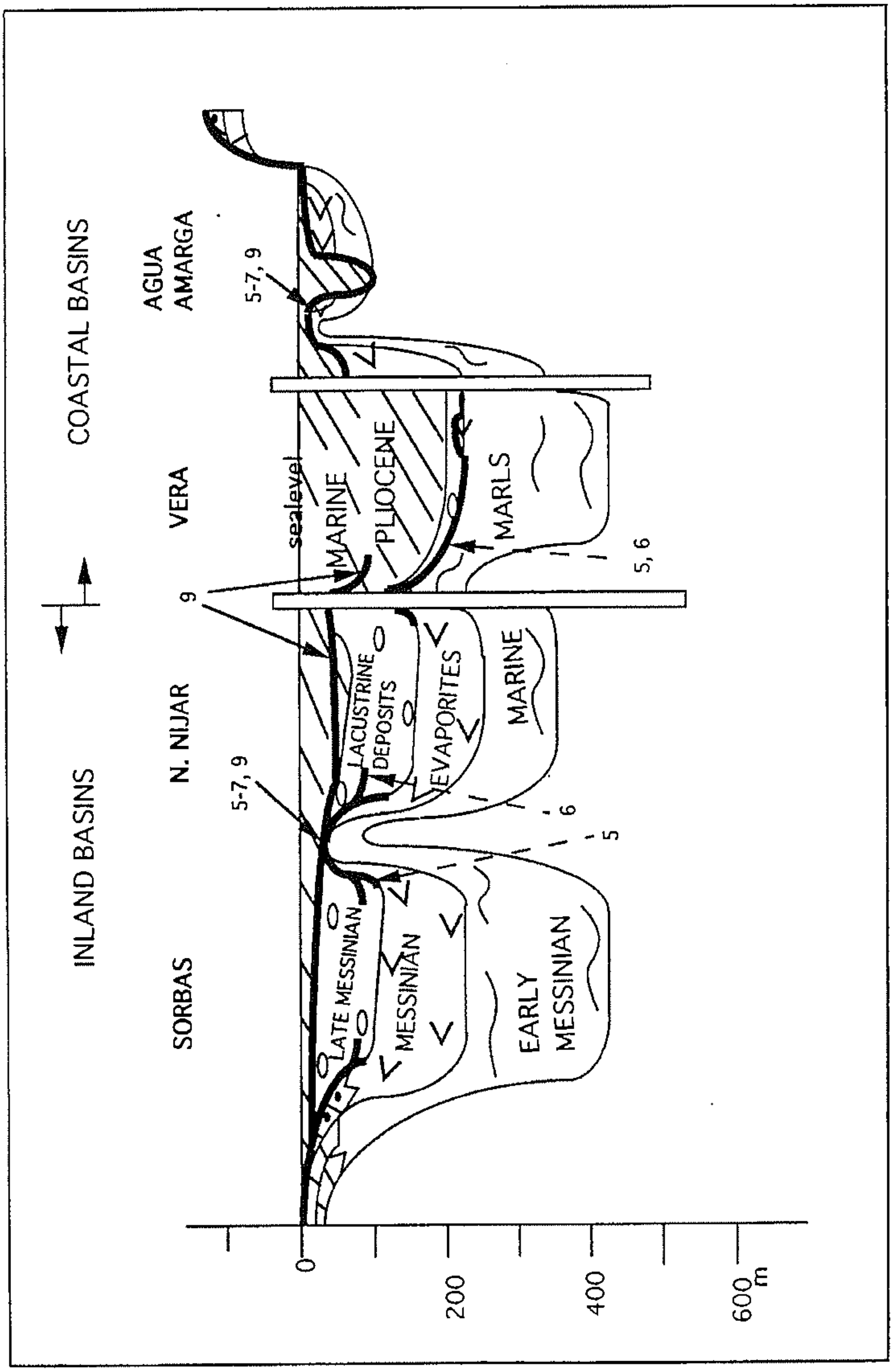
As the result of rapid evaporite deposition in the middle Messinian and the, localized, tectonic movements, late Messinian brackish lacustrine deposition took place in relatively shallow basins in the EAP. This holds especially for the less subsiding and well-filled Sorbas Basin and the tectonically elevated Agua Amarga area. In the Northern Nijar and Vera basins, deeper lacustrine deposition still could take place because they were still relatively subsident, probably in relation with N-S directed transcurrent movements, and since the latter was previously deeply eroded (Figs 5.8 & 5.9). A connection between the Northern Nijar and Sorbas basins still existed, but uplift of the Agua Amarga Basin and adjacent Castillico Ridge protected these basins from the direct erosive influence of Mediterranean dessications. There are presently no indications for a major barrier with the deeper Mediterranean to the SW of the Northern Nijar Basin, but the wide shelf formed by the Southern Nijar Basin and the Gulf of Almeria may have retarded erosive processes coming from this direction.

Only major tectonic subsidence can explain the sudden opening of a deep Gibraltar gate at the beginning of the Pliocene (Berggren & Haq 1976, Van Couvering et al. 1976, Müller & Hsü 1987, McKenzie & Sprovieri 1990). However, in our area, such subsidence is not recorded and the return of marine conditions probably initially remained limited to areas where creation of accomodation space was essentially the result of major late Messinian erosion (Fig. 5.9).

In our area, rather sedimentary infill and marginal uplift continued to be the prevalent processes, untill the area definitely emerged towards the end of the Pliocene (Figs 5.7-5.9). This was still followed by important differential uplift in relatively recent times (Fig. 2.3).

FIGURE 5.9. Schematic reconstructed NW-SE cross-section through main basinal areas and corridors of the EAP at the end of the Pliocene (with the Vera Basin 'artificially' displaced to the SE), showing differences in sedimentary fill and main unconformities near the Mio-Pliocene boundary (heavy lines). Numbers refer to facies changes (events) associated with the unconformities.

Note the important development of marine marls, Messinian evaporites and 'lago mare' deposits in the more protected inland areas, leading to loss in accomodation space, and the reduction of the former in the coastal basins bordering the 'Deep Mediterranean'. The marine Pliocene shows an inverse relationship filling wide incised valleys due to late Messinian Mediterranean 'sealevel' drawdown and to some extent to Mio-Pliocene uplift (as for the Agua Amarga Basin in particular , cf. Fig. 5.8). ------------------>


INLAND BASINS
COASTAL BASINS
SORBAS
N. NIJAR
VERA
AGUA AMARGA
sealevel
MARINE
PLIOCENE
MARLS
5, 6
9
5-7, 9
5-7, 9
LACUSTRINE DEPOSITS
EVAPORITES
MARINE
LATE MESSINIAN
MESSINIAN
EARLY MESSINIAN
6
5
0
200
400
600 m

CHAPTER 6

CONCLUSIONS

Environmental development in the Eastern Almeria Province

In the Eastern Almeria Province (EAP) deposition during the Messinian and in the very beginning of the Pliocene took place in a wide variety of (deep) marginal, and restricted continental basin situations. This episode is intercalated between periods of open marine sedimentation of oceanic type. Only slightly restricted "open sea" conditions existed in the earliest Messinian and earliest Pliocene, 'open lagoonal' conditions were widespread in the late early Messinian, whereas restricted, hypersaline marine conditions dominated in the middle Messinian. The late Messinian was characterized by a predominent (?exclusive) continental environment. Thus, a gradual increase in restriction of the marine environment and development to eventual continental conditions at the end of the Miocene is observed, whereas the return to open marine conditions in the Pliocene was relatively abrupt.

Environmental development of the EAP in the Mio-Pliocene boundary interval is further characterized by three other depositional features. The first is a net decrease in depositional depths between the Tortonian and the Pliocene. The second is an upwards decrease in warm-subtropical surface-water biota leading to changes in the planktonic foraminifer assemblages and the disappearance of typical subtropical carbonate platform facies in this interval. The development of consistent cyclic lithologic development, reflecting rhythmic variations in waterdepths, associated with variations in salinity, oxygen and nutrient content and temperature, is the third characteristic feature.

Environmental development is relatively uniform over the different basins of the EAP and it can be described in terms of a number of subsequent 'Environmental stages' and 'Substages':

'Stage Ia'. In the Tortonian and earlier Neogene, deep, subtropical, open oceanic conditions were predominant in the EAP. Biota diversity was high and authigenic mineral content essentially restricted to glauconite.

'Stage Ib'. In the earliest Messinian 'open sea' conditions without major oxygen-deficiency, nutrient excess or salinity variations are predominant and facies with only minor decrease in marine biota diversity and increase in authigenic mineral content develop

'Stage II'. In the late early Messinian deposition took place under moderately deep, 'open lagoon' conditions with important oxygen-deficiency at the sea-bottom, common nutrient-excess but still relatively minor variations in salinity of bottom and surface waters. The facies are characterized by low marine biota diversity, but relatively common high abundancy and a moderately high authigenic mineral content (dispersed organic material, sedimentary ore, carbonate, evaporite).

'Stage III'. During the middle Messinian, still moderately deep, '(semi)-restricted lagoon' conditions with common moderate to high salinity and also relatively common oxygen deficiency predominated and the environment was primarily characterized by abundant authigenic mineral formation (evaporites, carbonates, some sedimentary ore), whereas, except for bacteria, biota development was poor with extremely low diversity of the assemblages.

'Stage IV'. The late Messinian saw restricted lacustrine conditions, initially in a still predominantly hypersaline, later in a predominantly brackish environment. Facies rich in reworked clastic components but with a moderately high authigenic mineral content and relatively low continental biota abundancy and diversity are the most common. Such facies are best represented in the inland Northern Nijar and Sorbas basins, where lacustrine conditions were only periodically interrupted by fluviatile deposition, calcrete formation or subaerial erosion episodes in the central part of these basins. In the coastal Vera and Agua Amarga basins on the other hand, subaerial erosion and karstification of older Messinian deposits are the predominant late Messinian features, primarily ascribed to major drawdown of the adjacent 'Mediterranean Sea'.

'Stage V'. Still slightly restricted, 'open lagoon' to 'open sea'' conditions, with minor oxygen-deficiency, nutrient-excess and salinity deviations locally return in the earliest Pliocene, where late Messinian erosion had created "accomodation space". Where, as in the Sorbas Basin, and to a lesser extent, in Northern Nijar Basin, rapid filling with Messinian sediments was the predominant tecto-sedimentary process, or where, as in the Agua Amarga Basin uplift during the late Messinian-earliest Pliocene played an important role, fully open marine conditions, characterized by high marine fossil diversity, including typically 'oceanic' forms, and low authigenic mineral content became directly installed, but probably only after early Pliocene subsidence.

Implications for Mediterranean basin evolution

Our data confirm the importance of Mediterranean isolation at the end of the Miocene and the development of a 'Salinity Crisis'. Development of crisis conditions must have been gradual and its termination relatively abrupt. The Mediterranean as a whole was a deep marginal basin of slightly restricted 'open sea' type in the earliest Messinian. In the late early Messinian oxygen-deficiency became quasi-generalized and an 'open lagoon' came into existence. The nutrient budget of the basin was relatively high supporting a high marine organic productivity by a relatively limited number of organisms.

The middle Messinian sediments record predominant hypersaline conditions in a deep, (semi-)restricted lagoon with continuous inflow and high evaporation. Evaporite and other authigenic mineral production reached its maximum, whereas marine organic production was extremely low, again apart for that by bacteria. A period in the middle-late part of this interval saw better oceanic connections and outflow of at least the more superficial water, with occasionally resumed relatively high marine organic productivity, probably associated with a eustatic highstand. The major part of the up to two kilometers thick Messinian salt deposits of the central Mediterranean were probably deposited during a relatively short interval towards the end of this period and induced some shoaling of the deepest areas.

In the late Messinian, Mediterranean isolation became complete or virtually so. A variety of continental conditions developed with predominance of an evolving moderately hypersaline to more brackish brackish to "lago mare". At least one and probably a number of dessications of this enormous lake are documented by our data.

The environmental evolution recorded in the Vera Basin shows a relatively abrupt return to open marine conditions at the beginning of the Pliocene with a basin of 'open lagoon' to 'open sea' type, which rapidly became an 'open ocean'.

Factors in the basin evolution

Environmental development at both, the beginning of the Messinian and the beginning of the Pliocene indicate important variations in relative sill depth, which must primarily be ascribed to tectonic processes in the Entrance Area. Late Tortonian tectonic movements as recorded in our area first played an important role. However, subsequent Messinian tecto-sedimentary processes (uplift and infill) in the Entrance Area still appear to have been an important factor in the Mediterrranean isolation and its reduction in sill-depth.

The influence on the facies development of variations in global sealevel, in regional climate and of the Mediterranean basin configuration was relatively minor but their importance increased in the course of the Messinian, when sill depth had become sufficiently critical. Some global cooling over the Mio-Pliocene boundary interval probably also was a secondary factor. Regular, smaller variations in global climate with repercussions on both the sealevel and the regional water-budget played an important role in the cyclic development of the deposits.

The local geodynamic development

At the end of the Miocene the basins of SE Spain underwent important modification. Open marine sedimentation in deep, E-W elongated troughs which had a good connection with the open ocean was common during most of the Miocene. Towards the end of the Tortonian important structural rearrangement took place during a tectonic phase which entrained permanent changes. Shallower, more individualized and N-S directed basins came into existence which were more isolated from the open ocean as shallower parts of an enclosed Mediterranean basin during the Messinian.

The rapid infill with marine productive sediments in the early Messinian, with evaporites in its middle and with common erosional products in its upper part lead to further shoaling, shrinking and individualization of these basins aided by tectonic uplift of their margins. Relatively humid climatic periods are recorded at the end of the Messinian in our area, as in the rest of the Mediterranean, and resulted in widespread lake development. All the basins had become very shallow, but important erosion in the coastal areas created new 'accomodation space' for the subsequent Pliocene transgression.

Differential tectonic movements in the early Pliocene further reduced the marine area, but also locally enhanced the transgression of the Pliocene sea.

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