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1   Palynological analysis of Upper Ordovician to Lower Silurian sediments from the Diyarbakir

2   Basin, southeastern Turkey

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7   ABSTRACT

8   This paper reports on a palynological analysis of 41 core and 21 cutting samples from a well

9   drilled through an Upper Ordovician-Lower Silurian sequence, belonging to the Bedinan and

10   Dadas formations, of the Diyarbakır Basin of southeastern Turkey. The samples yield

11   abundant and well-preserved marine palynomorphs (acritarchs, chitinozoans and

12   scolecodonts) although non-marine palynomorphs (spores/cryptospores) are extremely rare or

13   absent. The Upper Ordovician sediments of the Bedinan Formation have low organic content

14   but contain abundant palynomorphs whereas the Lower Silurian sediments of the Dadas

15   Formation have high organic content, dominated by amorphous organic matter, with relatively

16   rare palynomorphs. Three chitinozoan assemblages are identified and attributed a late Katian

17   (merga Biozone), Hirnantian (moussegouda Biozone) and Llandovery (alargada Biozone)

18   age. Two acritarch assemblages are identified and attributed a Katian-Hirnantian and

19   Llandovery (Aeronian-Telychian) age. The chitinozoan and acritarch age determinations are

20   compatible and suggest that the Bedinan Formation is of Katian-Hirnantian age and is

21   separated by an unconformity from the Dadas Formation that is of Llandovery (Aeronian-

22   Telychian) age. These findings confirm the presence of an unconformity at the Ordovician-

23   Silurian transition in the southeastern Turkey. Palynofacies analysis suggests that the Bedinan

24   Formation accumulated on an offshore shelf that was initially well oxygenated but became

25   increasingly anoxic, whereas the Dadas Formation accumulated in an offshore basin that was

26   anoxic. Palynomorph assemblages recorded in the Bedinan and Dadas formations indicate

27   northern Gondwana affinity.

28   Keywords: Ordovician, Silurian, acritarchs, chitinozoans, Bedinan Formation, Dadas

29   Formation, Turkey.

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1. **Introduction**

This study examines chitinozoan and acritarch assemblages from Upper Ordovician and Lower Silurian sediments penetrated by a well in the Diyarbakır Basin, southeastern Turkey. The palynomorph assemblages comprise well preserved palynomorphs of low thermal maturity. This interval is of interest because other Lower Palaeozoic sequences of Northern Gondwana (Middle East, North Africa) are important targets for petroleum exploration. In general the lower Silurian organic-rich rocks ('hot shales') are excellent source rocks (Lüning et al., 2003) and glacial related rock sequences of the Upper Ordovician often form reservoirs (Le Heron et al., 2009).

The Diyarbakır Basin of southeastern Turkey is one of a number of early Palaeozoic basins on the North Gondwana platform. It includes a sequence mainly represented by siliciclastic sediments that are included in the Bedinan and Dadas formations. Unfortunately at present understanding of the subsurface geology of these deposits is poor and age dating of many of the sequences is insecure. Previous age assignments were mainly based on trilobites and graptolites (Paris et al., 2007). Macrofaunas were previously studied from the Taurus Range and Border Folds (Dean, 1967; Dean and Monod, 1990; Dean et al., 1999). Sarmiento et al. (1999) also described conodonts. However, extensive intervals of the succession are not well age-constrained because macrofaunas are sporadic in occurrence (Paris et al., 2007).

Very few palynological studies have been undertaken on the Lower Palaeozoic sequences of Turkey. Ordovician-Silurian acritarchs and chitinozoans have been described by Erkmen and Bozdogan (1979), Miller and Bozdogan (1989) and Steemans et al. (1996). Steemans et al. (1996) also reported on Ordovician and Silurian cryptospores and miospores. The Upper Silurian to Lower Devonian Dadas and Hazro formations were also studied palynologically by Fontaine et al. (1980) and Brocke et al. (2004). These studies may be considered alongside preliminary macrofossil data from both the Taurus Range and southeastern Turkey (Dean and Martin, 1992; Dean et al., 1993; Monod et al., 2003, Paris et al., 2004; Sancay and Dinc, 2012).

A detailed palynological analysis is desirable for: (i) determining the biostratigraphical age of the strata and improving stratigraphical correlation; (ii) enabling interpretation of palynofacies and palaeoenvironments. The main aim of this study is to present a preliminary palynological investigation of these deposits based on material from core and cutting samples from a well drilled through the Dadas and Bedinan formations.

2. Geological setting

The history of research on recognizing Palaeozoic rocks in Turkey begins in the second half of the 19th century when a large number of scientists began working on the stratigraphy and palaeontology of the Palaeozoic rocks of this region (e.g. Tchihatcheff, 1864; de Verneuil, 1869). The systematic mapping of these Palaeozoic rocks was primarily conducted by the Mineral Research and Exploration Institute of Turkey and the Turkish Petroleum Company, but also by some independent geologists (e.g. Tolun and Ternek, 1952; Ketin, 1966; Dean, 1967 to 2006; Kaya, 1973).

Turkey is composed of a number of continental fragments which were separated by ancient oceans until the Cenozoic. The continental fragments (terranes) and their remnants, such as ophiolites and accretionary prisms, subsequently came together to form Anatolia (Okay, 2008). Goncuoglu (2012) notes that Turkey is located in the centre of the Alpine orogenic belt and was formed by closure of at least three branches of the Neotethys Ocean between Laurasia in the north and Gondwana in the south.

Ketin (1966) divided Turkey into three main tectonic terranes (the Pontides, Anatolide-Tauride and Arabian Platform), but recently new findings have emended the classification of these continental fragments. It is now considered that Turkey consists of five main terranes connected by different suture belts. In northern Turkey there are three terranes: the Rhodope-Strandja, İstanbul-Zonguldak and Sakarya Terranes (together called the Pontides). In southern Turkey the Anatolite-Tauride terrane is separated from the Pontides by the Izmir-Ankara-Erzincan suture zone (Okay, 2008; Goncuoglu, 2012). The Pontides has Laurasian affinities while the Anatolite-Tauride terrane has Gondwana affinities. It is currently a controversial issue as to whether the Central Anatolian Crystalline Complex or the Kırsehir Massif is part of the Anatolite-Tauride terrane or an independent terrane (Okay, 2008). The Southeast Anatolian Autochthon is a northern continuation of the Arabian platform and resembles the Anatolite-Tauride terrane stratigraphically (Okay, 2008; Goncuoglu, 2012).

The Southeast Anatolian Zone in the south and the Bitlis Unit in the north are separated from each other by an active thrust that is a continuation of the Zagros fold and thrust belt (Goncuoglu, 2010). Pan-African basement and Palaeozoic–Cenozoic cover are the main components of the Southeast Anatolian Zone (Sungurlu, 1974). The basement complex is covered by arenite which is intercalated with shelf carbonates and Mid Cambrian aged

95   nodular limestone (Schmidt, 1966; Sungurlu, 1974; Dean, 2006; Demircan and Gursu, 2009).

96   Ordovician siliciclastic rocks are observed below the Upper Silurian-Upper Devonian

97   succession with a depositional break occurring during the Early Silurian. Late Silurian-Late

98   Devonian deposits are composed of continental clastics and limited marine sediments in the

99   central part (Goncuoglu, 2010). Perincek et al. (1991), on the other hand, considers that the

100   Ordovician rocks are covered by Upper Devonian-Lower Carboniferous coastal-shallow

101   marine rocks in the eastern areas. The northern part of Gondwana was stable during the Late

102   Palaeozoic resulting in intercalations of two units: Late Permian shelf carbonates and Triassic

103   shallow marine deposits (Goncuoglu, 2010).

104   Lower Palaeozoic sedimentary rocks of southeastern Turkey are observed in many

105   localities (Fig. 1). Southeastern Anatolia is located at the northern end of the Arabian Plate

106   (Monod et al., 2003) and the tectonic evolution and sedimentary sequence of the region are

107   controlled by the relative movements of the Arabian Plate and the Anatolian Continent with

108   respect to one another. The Mardin-Kahta High, which was formed by the influence of these

109   movements, separates Southeastern Anatolia into two sub-basins (the Akçakale Basin and

110   Diyarbakır Basin) (Bozdogan and Erten, 1990).

111   The studied well is located in the province of Diyarbakır in the Diyarbakır Basin

112   (Fig.1). The formations sampled through the well are the Bedinan and Dadas formations (Fig.

113   2 and 3).

114   2.1. Bedinan Formation

115   Steemans et al. (1996) stated that the thickness of this formation is 1500 m and the

116   depositional environment is shallow marine and tidal-flat. Transgression is observed in the

117   lower part of the formation whereas two successive regressions are observed in the upper part.

118   The upper part of the sequence is comprised of dark green shales and sandstone (Bozdogan et

119   al., 1994). According to Bozdogan et al. (1996) the lower part is composed of fossiliferous

120   dark shales and siltstone, and the middle and upper parts are composed of sandstones and

121   shales with local submarine lavas.

122   The Bedinan Formation is recognised at outcrop and in subsurface boreholes. It yields a

123   rich benthic fauna and palynomorphs (Bozdogan et al., 1994; Steemans et al., 1996). The

124   lower part of the formation is Sandbian (Carodoc) while the upper part is Katian-Hirnatian

125   (late Caradoc-Ashgill) (Dean et al., 1981). The Bedinan Formation is unconformably overlain

126 by the Dadas Formation and in some regions stratigraphical differentiation is apparent
127 because of erosion that occurred after the Ordovician period (Steemans et al., 1996).

128   2.2. Dadas Formation

129   The thickness of the Dadas Formation varies from 100 to 400 m in the subsurface and it
130 is exposed in Korudag and Dadas (Diyarbakir region) (Steemans et al., 1996). The formation
131 is underlain by the Bedinan Formation and overlain by the Kayayolu and Hazro formations
132 (Bozdogan et al., 1987). Ozdemir (2013) suggests that the Dadas formation is conformably
133 overlaid by the Hazro Formation, and the lower boundary of the Dadas Formation is not
134 exposed at the surface. However, it unconformably overlies the Bedinan Formation in the
135 Kayayolu-2 well. The lower part of the Dadas Formation is composed overwhelmingly of
136 anoxic organic-rich shale and limestone whereas the upper part of the formation is composed
137 of sandstone and dolomite, which were deposited under high energy conditions in a shallow
138 marine environment (Steemans et al., 1996). Bozdogan et al. (1996) suggests that the lower
139 part of the formation represents an inner-shelf with low energy condition. Shallow marine
140 deposits in the middle part and coastal deposits in the upper part are intercalated. Previous
141 palynological studies suggest a mid Silurian to Early Devonian age for the formation
142 (Bozdogan et al., 1987, 1994).

143

144   **3. Materials and methods**

145   Sixteen core and twenty-one cutting samples from the Upper Ordovician rocks and
146 twenty-five core samples from the Lower Silurian rocks were selected. The core from the
147 depth of 1808 to 1873.5 m penetrated 33.95 m of the Dadas Formation and 12.4 m of the
148 underlying Bedinan Formation. Core samples were derived from the interval 1826.05-1873.4
149 m and cuttings were derived from the interval 1908-2362 m.

150   Samples were prepared in the palynology laboratory of the Turkish Petroleum
151 Research and Development Centre. Standard HCI-HF-HCI palynological acid maceration
152 techniques were applied prior to heavy liquid separation. No oxidation was needed. Samples
153 generally yielded assemblages of abundant and well-preserved palynomorphs.

154   The prepared slides were examined using an OLYMPUS BH-2 microscope with
155 specimen location recorded using an England Finder. Photographs of the specimens were

156   obtained using a MEIJI, transmitted light microscope, with a digital camera that uses Infinity

157   Analyze software. Counts for palynofacies analysis considered only palynodebris larger than

158   10 µm in diameter. All material (rock sample, residue, slides and digital images) are curated

159   in the palynology laboratory of the Turkish Petroleum Research and Development Centre.

160

161   4.   Biostratigraphy and age

162   4.1. Chitinozoan Assemblages

163   Three chitinozoan assemblages (CA1 to CA3) have been defined within the interval

164   2362 to 1826.05 m (Figs 4 and 5). In the interval of 1930-2362 m the oldest Chitinozoan

165   Assemblage 1 is observed and is assigned a late Katian age. The overlying Chitinozoan

166   Assemblage 2 from 1870.3-1912 m is assigned a Hirnantian age. The youngest Chitinozoan

167   Assemblage 3 is identified in the sequence from 1826.05 to 1857.5 m and is assigned a

168   Llandovery (Aeronian-Telychian) age.

169   4.1.1. Chitinozoan Assemblage 1 (1930-2362 m)

170   CA1 is assigned to the merga Biozone. It is characterized by the total range of

171   Ancyrochitina merga as well as the diversity and abundances of Ancyrochitininae (Paris,

172   1990).   Armoricochitina   nigerica,   Calpichitina   lenticularis,   Conochitina   dolosa,

173   Cingulachitina sp., Cyathochitina sp. and Desmochitina minor are present, and Plectochitina

174   cf. sylvanica appears with Rhabdochitina magna. The upper boundary of CA1 is identified by

175   the sharp disappearances of A. merga and other Ancyrochitininae species. This may

176   correspond to the first phase of the end-Ordovician (Hirnantian) extinction at the base of

177   extraordinarius graptolite biozone (Harper et al., 2013). The lower boundary of CA1, on the

178   other hand, cannot be determined because the last downhole occurrence of A. merga is not

179   observed in these samples. However, an early nigerica Biozone is likely from the depth of

180   2162 m downward if we consider the possibility that caving is responsible for the presence of

181   A. merga in the samples from 2184-2186, 2352-2354 and 2360-2362 m (Fig. 4). There is clear

182   evidence for caving in some of the cuttings samples which complicates biostratigraphical

183   interpretation. For example, from the depth of 1908 m downward caved Silurian chitinozoans

184   (Angochitina macclurei) and acritarchs (Cymbosphaeridium pilar, Cymbosphaeridium sp. 1,

185   Dictyotidium dictyotum, Deflandrastum millepiedi) are present in cutting samples (Figs 4 and

186   6). Consequently, based on potential caving within CA1, the merga and nigerica Biozones are

187   not confidently separated in this well. Chitinozoan species like Hyalochitina fistulosa and

188   Belonechitina robusta which have their FAD earlier (see Paris et al., 2007 and Paris, 1990)

189   have not been recorded in the studied interval. Thus there is no age control to identify an age

190   older than late Katian and consequently this zone is assigned to the late Katian (Fig. 5).

191

192   4.1.2. Chitinozoan Assemblage 2 (1870.9-1912 m)

193   CA2 is similar to CA1 and is characterized by typical Late Ordovician species

194   including Armoricochitina nigerica, Calpichitina lenticularis, Conochitina dolosa,

195   Euconochitina cf. moussegoudaensis and Desmochitina minor. However, it differs from the

196   underlying biozone CA1 due to the absence of A. merga. This zone can be assigned to the

197   moussegoudaensis/oulebsiri Biozone based on the occurrence of Euconochitina

198   moussegoudaensis (Thusu et al., 2013). These authors suggested that the moussegoudaensis

199   Biozone starts with the FAD of E. moussegoudaensis with this taxon possibly ranging through

200   the Ordovician-Silurian boundary in southeast Libya. However, E. cf. moussegoudaensis

201   identified in this study ranges through this zone as well as CA1. Based on the discussion

202   above, and the uncertainty of the age assignment of E. moussegoudaensis, this zone is

203   cautiously assigned a Hirnantian age (as discussed by Thusu et al. 2013) (Fig. 5).

204   Diagnostic Hirnantian species (e.g. Spinachitina oulebsiri and Tanuchitina elongata)

205   are absent in this assemblage, which may be a consequence of unfavourable environments for

206   Hirnantian taxa. During the second phase of the end-Ordovician mass extinction the climate

207   drastically changed from cooling to warming, causing the melting of major ice caps, sea level

208   rise and widespread anoxia (Harper et al., 2013). Melt water and invasion of continental

209   shelves by anoxic bottom waters during this transgression may have created unfavourable

210   brackish and/or anoxic environments excluding certain Hirnantian chitinozoan taxa (as

211   discussed by Melvin, 2014).

212   4.1.3. Chitinozoan Assemblage 3 (1826.05-1857.5 m)

213   CA3 is characterized by the occurrence of Angochitina sp., Bursachitina sp.,

214   Lagenochitina navicula, Pterochitina deichaii and Conochitina cf. alargada. It is cautiously

215   assigned to the alargada Biozone in which C. alargada (=C. edjelensis alargada) is

216   dominant. This biozone is defined by the concurrent range of C. alargada and Plectochitina

217   paraguayensis (Paris et al., 1995). Verniers et al. (1995) showed that the age range of C.

218   alargada extends from the mid to late Aeronian. Paris et al. (2015) suggest that the alargada

219   Biozone is found in the early Aeronian sediments of Arabia, although specimens referred to as

220   C. cf. alargada extend up into the Telychian in this region (Paris et al. 2015). One of the

221   accompanying species in CA3 is P. deichai that is present in low numbers in this Biozone.

222   Another accompanying chitinozoan in this assemblage is L. navicula, which was reported

223   from the Llandovery of northeast Brazil by Grahn et al. (2005). It is worth emphasising that

224   Angochitina macclurei has been found as caving within the 1930-2290 m interval. The

225   occurrences of these taxa suggests that the macclurei (early Telychian) Biozone is likely

226   present within or above the cored interval (see Paris et al., 2015). Based on the discussion

227   above and the occurrence of C. cf. alargada this assemblage is assigned an Aeronian-

228   Telychian age (Fig. 5).

229

230   4.2. Acritarch Assemblages

231   Two acritarch assemblages have been identified (AA1-AA2). Acritarch Assemblage 1 is

232   older and recovered from the Bedinan Formation and Acritarch Assemblage 2 is younger and

233   recovered from the Dadas Formation (Fig. 6).

234   4.2.1. Acritarch Assemblage 1

235   This assemblage is mainly composed of Villosacapsula setosapellicula,

236   Orthosphaeridium bispinosum, Orthosphaeridium chondrododora, Multiplicisphaeridium

237   irregulare, Orthosphaeridum quadrinatum, Ordovicidium elengatulum, Orthosphaeridum

238   ternatum, Dactylofusa striata, Dactylofusa platynetrella, Leiofusa litotes, Baltisphaeridium cf.

239   latiradiatum, Baltisphaeridium longispinosum, Veryhachium subglobosum and Veryhachium

240   lairdi (Plates 4-8).

241   O. bispinosum, O. chondrododora, O. quadrinatum B. cf. latiradiatum were

242   previously reported from Caradoc (Sandbian-Katian) sediments of Britain (Turner, 1979), but

243   these taxa can also extend into the Ashgill (Katian-Hirnantian) (Vecoli and Le Herisse, 2004,

244   Le Herisse et al., 2015). Although Dactylofusa striatogranulata (Plate VI, 9) can extend from

245   the Late Ordovician into the Lower Silurian (see Jardiné et al., 1974, Le Herisse et al., 2015)

246   it was restricted to the uppermost Ordovician (1870.9 m herein) in the study of Vecoli and Le

247   Herisse (2004). However, Dactylofusa platynetrella (Plate V, 9) is not only restricted to the

248   Hirnantian sediments (Vecoli and Le Herisse, 2004). It can be found in the Katian sediments

249   recovered from the well (1870.9-2042 m). D. platynetrella were also recorded from the

250 Katian and Hirnantian sequences of northern Iran with occurrences of Villosacapsule
251 setosapellicula, Orthosphaeridium elengatulum, Dactylofusa striata (Ghavidel-Syooki et al.,
252 2011).

253 V. subglobosum, V. lairdi and V. setosapellicula were previously documented both
254 from the Sarah and Qasim formations of the central Saudi Arabia (Le Herisse et al., 2015).
255 These taxa are observed in the Hirnantian and Katian parts of the Bedinan Formation. V.
256 subglobosum and V. setosapellicula in the Hirnantian sediments are distinctly higher than
257 those in the Katian part of the Bedinan Formation. O. chondrododora, M. irregulare, O.
258 quadrinatum, O. elengatulum, O. ternatum and V. setosapellicula were reported from the
259 early mid Katian sediments of Oman (Droste, 1997). However, M. irregulare, O.
260 quadrinatum, O. elengatulum are restricted to the Hirnantian in this study (Fig. 6). Previously,
261 many Late Ordovician indicative acritarchs were recorded from the southeastern part of
262 Turkey (Paris et al., 2007; Steemans et al., 1996). Furthermore, L. litotes (Plate V, 10)
263 documented from the Caradoc-Ashgill (Katian) boundary of northeast Kansas (Wright and
264 Meyers, 1981) is observed here through the late Katian-Hirnantian of the Bedinan Formation.

265 Peteinosphaeridium nudum (Plate VII, 10) ranges through the Lower Ordovician and
266 possibly Middle Ordovician rocks of northern Gondwana (Vecoli and Le Herisse, 2004) but it
267 is observed at 1872.3 m, indicating probable reworking. Peteinosphaeridium trifurcatum is
268 another long-ranging Ordovician acritarch from northern Gondwana (Vecoli and Le Herisse,
269 2004). Peteinosphaeridium trifurcatum subsp. trifurcatum (1872.8 m, Plate VII, 8) and
270 Baltisphaerosum christoferii (1872.8 m, Plate VII, 12) is likely reworked here since these
271 species were recorded from the Landeilo and Caradoc (Darriwilian-Katian) of Britain (Turner,
272 1979). Baltisphaeridium annelieae was documented from the Llandovery and lower Wenlock
273 of Wales and the Welsh Borderland of Britain (Hill, 1974). However, in this study,
274 Baltisphaeridium annelieae? (1870.9 m, Plate IV, 11) is observed in the Bedinan Formation.
275 The range of this species is still debatable.

276 AA1 includes Silurian caved acritarchs such as Cymbosphaeridium pilar,
277 Cymbosphaeridium sp., Dictyotidium dictyotum, Onondagella sp., Deflandrastum millepiedi,
278 and Pirea sp.

279 Based on the above discussion, particularly regarding classical Late Ordovician
280 (Sandbian-Hirnantian) acritarchs such as Baltisphaeridium longispinosum, Baltisphaerosum
281 onniensis, Veryhachium subglobosum, Veryhachium oklahomense, Villosacapsula

282   setosapellicula and Leiofusa litotes, AA1 and the Bedinan Formation is assigned to the Late
283   Ordovician. LADs of Stellechinatum celestum, or earlier Frankea spp. and Dicrodiacrodinium
284   ancoriforme, are not observed through the well. Therefore, a late Katian-Hirnantian age
285   appears most likely rather than Sandbian-early Katian.

286   4.2.2. Acritarch Assemblage 2

287   AA2 is characterized by the existences of Multiplicisphaeridium fisheri, Oppilatala
288   eoplanktonica, Veryhachium wenlockium, Veryhachium europaeum, Domasia trispinosa,
289   Dactylofusa striatifera, Visbysphaera dilatispinosa, Visbysphaera oligofurgata and
290   Multiplicisphaeridium ramusculosum (Plates 1-3). This suggests an Aeronian-Telychian age
291   when considered together with the chitinozoan age designation. M. fisheri and O.
292   eoplanktonica, reported from the early Silurian shales of Libya (Paris et al., 2012), indicates a
293   Rhuddanian-Aeronian age. M. fisheri (1829-1857 m in this study), on the other hand, is found
294   from the earliest Telychian to the latest Sheinwoodian (Loydell et al., 2013). The Aeronian-
295   Telychian age interval is also consistent with the existence of D. striatifera (or Eupoikilofusa
296   striatifera), which indicates an age no older than Rhuddanian (Paris et al., 2012). However,
297   Loydell et al. (2013) suggested an extended age range (from Late Ordovician to earliest
298   Devonian) for Eupoikilofusa striatifera (1857.5 and 1867.5 m in the Dadas and Bedinan
299   formations respectively). In addition, occurrences of D. striatifera and V. europeoum are
300   coincident with a Llandovery age (Thusu et al., 2013). Furthermore, occurrences of V.
301   dilatispinosa, V. oligofurgata, M. ramusculosum,and V. wenlockiumrecorded in the Dadas-I
302   Member suggest a Llandovery age (Hill, 1974).

303   D. trispinosa (1855-1847 m herein) was previously reported from the early Silurian of
304   Iran and stratigraphic age range of D. trispinosa is known from the late Llandovery
305   (Telychian) to early Ludlow (Ghavidel-Syooki et al., 2011). Assemblage of D. trispinosa with
306   D. bispinosa is no older than the Telychian (Hill and Dorning, 1984). D. bispinosa was also
307   previously recorded from a similar palaeogeographic region in the late Llandovery (Keegan et
308   al., 1990). Moreover Wauthoz (2005) documented other Domasia species (D. elongata/D.
309   limaciformis) attributed to the Aeronian-Telychian boundary. Based on this discussion and
310   occurrences of D. trispinosa and D. bispinosa together at 1854 m the lower boundary of the
311   Telychian can be attributed to this level.

312

5. Palaeoenvironmental interpretation

Palaeoenvironmental interpretation of the studied intervals is based on palynofacies analysis (based on 200 counts of palynomorphs and palynodebris).

The lower part of the Bedinan Formation has low organic yield but contains a high abundance and diversity of both acritarchs and chitinozoans. Land-derived spores/cryptospores are extremely rare. Amorphous Organic Matter (AOM) is present but not abundant (Table 1). The depositional environment is interpreted as an offshore, open marine, shallow shelf that was oxygenated. The presence of abundant animal remains supports this interpretation. The upper part of the Bedinan Formation has lower palynomorph abundance (but still yields acritarchs and chitinozoans with only very rare spores/cryptospores) and higher amounts of AOM (Table 1). The depositional environment is interpreted as an offshore, open marine, deeper water shelf that was anoxic. Changes in AOM amount in this formation may corresponds to the study of Harper et al. (2013) who suggested that the end Ordovician (Hirnantian) extinction comprised two discrete pulses with the second phase of this extinction linked to a remarkable transgression resulted from melting of ice caps following global warming. Steemans et al. (1996) suggested that the depositional environment of the Bedinan Formation was shallow marine to tidal-flat environment based on the sections they studied that contained relatively common land-derived spores/cryptospores. The studied well, however, is located in the eastern part of the basin and yields only extremely rare non-marine palynomorphs.

The Dadas Formation has high organic yield and is dominated by orange to brown AOM (up to 95%). However, palynomorphs are abundant and include relatively common sphaeromorphs, rare to relatively common acritarchs and chitinozoans, although spores/cryptospores are extremely rare (Table 2). This suggests that the deposits accumulated in an offshore, open marine, deep basinal environment that was anoxic. The presence of abundant pyrite associated with the AOM supports interpretation of anoxic conditions. An environment that is dominantly represented by AOM almost certainly reflects anoxic conditions, especially in areas far from the effect of terrestrial input (Batten, 1996). According to Steemans et al. (1996) the lower part of Dadas Formation consists of anoxic organic-rich shale and limestone, which is consistent with the interpretation herein.

6. Conclusions

Detailed palynological analysis of the Upper Ordovician to Lower Silurian sediments from a well in the Diyarbakır Basin, southeastern Turkey has enabled biostratigraphical age determination and palaeoenvironmental interpretation based on palynofacies analysis.

Three chitinozoan assemblages and two acritarch assemblages have been identified. Biostratigraphical evidence from both chitinozoans and acritarchs are compatible but due to a higher precision of taxon ranges the chitinozoan age determinations are considered more accurate than those based on acritarchs. Thus the Bedinan and Dadas formations are considered to be Katian-Hirnantian and Llandovery (Aeronian-Telychian) in age, respectively, and separated by an unconformity.

Palynofacies analysis suggests that the Bedinan Formation accumulated on an offshore, high productivity shelf that became increasingly less oxygenated, whereas the Dadas Formation accumulated in an offshore, anoxic, low productivity basin. The remarkable deepening following melting of ice cover during the second phase of end-Ordovician mass extinction may be responsible for the increasingly anaerobic conditions in the Bedinan Formation.

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