

# Mineralogy and Characteristics of Soils Developed on Persian Gulf and Oman Sea Basin, Southern Iran: Implications for Soil Evolution in Relation to Sedimentary Parent Material

Hakime Abbaslou,<sup>1</sup> Ali Abtahi,<sup>2</sup> Francisco Jose Martin Peinado,<sup>3</sup> Hamidreza Owliaie,<sup>4</sup>  
and Farhad Khormali<sup>5</sup>

**Abstract:** There are relatively few comprehensive documents about the mineralogy of arid regions, and no study has been reported on the characteristics and mineralogical trends of the soils of arid regions in the Hormozgan Province, southern Iran. The objective of this study was to identify the variations in physical, chemical, and mineralogical composition of soils of the region lying between the southern part of Zagros and Persian Gulf to Oman Sea. X-ray diffraction, thin-section studies, and transmission electron microscopy with energy-dispersive X-ray spectroscopy were conducted to determine the relationship between mineral distribution and climate, along with an evaporate environment reflected by a variety of geological structures. On the basis of nature and composition, the soils can be classified into Alluvial soils rich in silt and Desert soils, sandy soils poor in organic carbon in which evaporate Tertiary Formations played an important role on the formation of contemporary minerals. The maximum portions of silt and sand fraction were composed of quartz and alkali-feldspars, whereas calcite and dolomite were elevated in soil samples without acid treatments. The clay mineralogy results revealed that detrital input and inheritance are possibly the main source of kaolinite, smectite, chlorite, and illite, whereas *in situ* neoformation during the Tertiary shallow saline and alkaline environment could be the dominant cause of palygorskite occurrences in the sedimentary rocks. The chemical composition of clay suspension, determined by electron microprobe analysis, corresponds to that of Mg-rich predominant palygorskite, with a degree of tetrahedral substitution from almost less than 2 to 23%. In conclusion, the soil mineralogy did not vary systematically with depth but showed spatial variations, and the relative influence of mineral distribution in topsoils can be summarized as follows: parent rocks > geomorphological setting > water table position > climate > physiography and soil evolution. The soils, in general, contained moderate to high amounts of weatherable minerals, indicating their early to intermediate stages of development.

**Key Words:** Arid soils, soil mineralogy, silt particles, palygorskite, thin section, Hormozgan Province  
(*Soil Sci* 2013;178: 568–584)

<sup>1</sup>Department of Civil Engineering, Sirjan University of Technology, Sirjan, Iran.

<sup>2</sup>Department of Soil Science, College of Agriculture, Shiraz University, Shiraz, Iran.

<sup>3</sup>Department of Soil Science, College of Sciences, University of Granada, Granada, Spain.

<sup>4</sup>Department of Soil Science, Faculty of Agriculture, Yasouj University, Yasouj, Iran.

<sup>5</sup>Department of Soil Science, Faculty of Water and Soil Engineering, Gorgan University of Agricultural Sciences and Natural Resources, Gorgan, Iran.

Address for correspondence: Dr. Hakime Abbaslou, Department of Civil Engineering, Sirjan University of Technology, Sirjan, Iran. E-mail: abbaslou@sirjantech.ac.ir

Received July 17, 2013.

Accepted for publication November 19, 2013.

Copyright © 2014 by Lippincott Williams & Wilkins

ISSN: 0038-075X

DOI: 10.1097/SS.0000000000000022

Iran, situated in the western part of Asia, is located in the mid-latitude belt of the arid and semiarid regions (UNESCO, 1977). About 65% of its total area is characterized by an arid climate (Banaei et al., 2005). Hormozgan Province, one of the most arid parts of Iran, covers an area of 70,138 km<sup>2</sup> located in the south of Iran and north of the Persian Gulf and Oman Sea. Because of its warm and dry climate as well as the limited rainfall, with exception of some rare elevations and mountainous regions, aridity is common throughout the region. However, soil reports for the Oman Sea and Persian Gulf margins are not available, at least in the open literature.

To address society's needs related to soil resources, it is important to understand the landscape distribution of soil minerals and the processes responsible for their occurrence. Soil minerals can be used to help understand soil pedogenesis (Owliaie et al., 2006; Graham and O'Geen, 2010) and the current behavior of wetlands (O'Geen et al., 2008) and arid lands (Reid et al., 1993); interpret paleoenvironmental conditions (Amundson et al., 1989; Khormali et al., 2003), inherent soil fertility issues (Page et al., 1967; Murashkina et al., 2007), soil water behavior and irrigation management (Alon et al., 2006), wind erosion and inhalation problems (Reheis, 2006), and soil aggregation and carbon sequestration potential (Rasmussen et al., 2005); and assist in forensic investigations (Stam, 2004).

The mineralogical compositions of the bulk soil, rock, and various grain size fractions are important soil formation factors that contribute to the distinction of provenance. The contribution of bed rock types to soil clay mineralogy is dependent on the intensity of weathering, which differs from one environment to another, and factors like relief (Johnsson, 1993), climate (Ruffell et al., 2002), and vegetation, and different bedrocks react differently to chemical weathering, resulting in various landscapes and weathering products (Pye, 1986; Twidale, 1990; Dultz, 2000).

It is a common belief that the mineralogy of Aridisols is largely derived from the parent material (Allen and Hajek, 1989; Nettleton and Peterson, 1983). Woodruff et al. (2009) studied two transects across the United States and Canada and revealed that, along both transects, soil mineralogy changed abruptly with changes in soil parent materials. However, the chemical influence of a soil's parent material can be obscured by changing climatic conditions. Soils that are texturally and chemically similar may still differ because of the presence or absence of small amounts of particular minerals (Van der Merwe et al., 2002). Aridisols in the Hisma Basin, southern Jordan, reveal a suite of minerals—calcite, gypsum, 2:1 clay minerals, K-feldspar, plagioclase, and dolomite, and possibly goethite, palygorskite, and chlorite—in the soil that contrasts markedly with the nearly monomineralic composition of the bedrock—quartz, with minor kaolinite and traces of illite (Ugolini et al., 2008). Quartz, K-feldspar, and muscovite are more resistant to chemical weathering than plagioclase and iron-bearing minerals such as biotite, chlorite, and amphibole

(Brady, 1990; Berner and Berner, 1996; Taylor and Eggleton, 2001). The regional grain size characteristics of eolian sediments from the Strzelecki and Tirari Deserts (Australia), combined with local occurrences of rock fragments and high feldspar content, suggest that the sediments are mostly derived from local sources (Fitzsimmons et al., 2009).

Various studies conducted on the clay minerals in the arid and semiarid environment of Iran showed the existence of chlorite, illite, smectite, vermiculite, and palygorskite minerals (Gharaee and Mahjoory, 1984; Khademi and Mermut, 1998; Khormali and Abtahi, 2003; Owliaie et al., 2006). However, there are no published reports on detailed mineralogy of the fractions of each soil separate (sand, silt, and clay) and associated rocks in arid regions of southern Iran. Also, the diversity of parent materials in southern Iran offers a good environment to study the distribution and origin of minerals in an arid climate.

Therefore, the purpose of this study was to compare the mineralogy of sedimentary rocks with that of the soil to establish the contribution made by sedimentary material to arid soils developed on evaporates and alluvium materials to investigate the mineralogy of different grain size particles (sand, silt, and clay) in arid regions with light textures to distinguish provenance and resultant factors on distribution and types of minerals. An additional purpose was to provide, for the first time, a detailed mineralogical analysis of soils and sedimentary rocks of Hormozgan Province, Iran.

## MATERIALS AND METHODS

### Study Area

Hormozgan Province is located in southern Iran (latitude 25° 24' to 28°57' N and longitude 52°41' to 59°15'E) with an area of

70,138 km<sup>2</sup> (Fig. 1). The major portion of the province has a mountainous terrain (36%), with the southern sector of the Zagros Mountain Range and Bashagard Mountain Range. The Zagros Mountain Range extends from northeast to southeast of the province and terminates at lime and sandy hills and highlands as well as coastal lowlands parallel to the Persian Gulf and Oman Sea. The Bashagard Mountain Range, belonging to the second and third geological periods, lies in the eastern region.

### Climate

The target area is situated in a desert to a semidesert-like region with a warm (hot) and dry climate. Rainfall is scant (185 mm as annual average), but humidity along the coastal regions is of a high percentage (Choopani et al., 2006). The soil temperature regime is hyperthermic, but some rare north and coastal areas have thermic and isothermic regimes, respectively (Fig. 1D). According to Newhall Simulation and subdivisions implemented by Van Wambeke (2000), the study region can be subdivided into Extreme Aridic, Aridic, Typic tempustic, and Xeric Tempustic soil moisture regimes (Fig. 1C). Overall, Aridic and Ustic soil moisture regimes were obtained for the whole Hormozgan Province. Table 1 shows the climatological data and P/ET<sup>o</sup> (ratio of mean annual precipitation to mean annual reference crop evapotranspiration) of the most important weather stations in Hormozgan Province.

### Geological Setting

The territory of this province is a junction of three important structural zones of Iran: Zagros, Makran, and Central Iran. Therefore, this province can be divided geologically into three distinct sectors: (i) Zagros zone, which is extending from northwest to

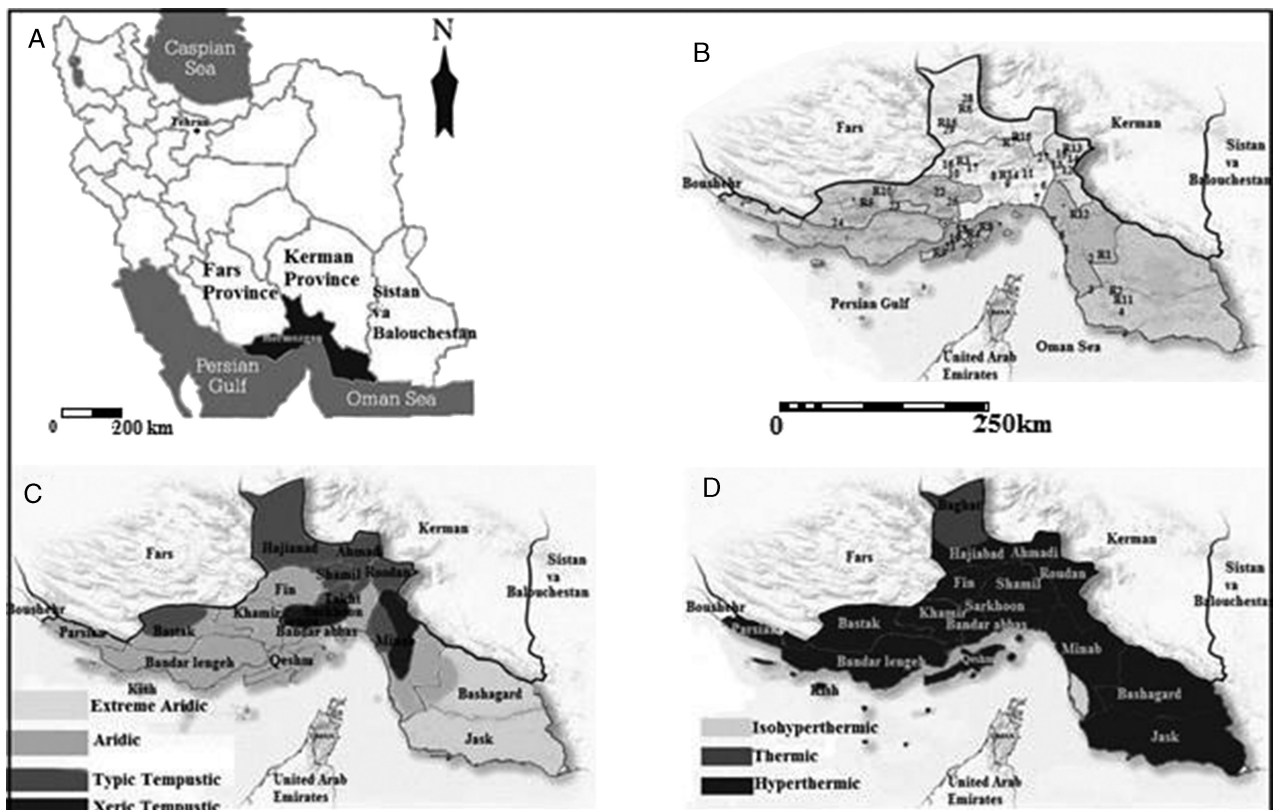


FIG. 1. (A) Location of Hormozgan Province in map of Iran. (B) Sites of studied pedons and rocks. (C) Soil moisture regime map. (D) Soil temperature regime map. No scale bar for the maps.

**TABLE 1.** Climatological Data of the Most Important Weather Stations in Hormozgan Province

| Weather Station Name   | Annual Temperature Average, °C | Annual Rainfall Average, mm | Climate   | P/ET <sup>o*</sup> |
|------------------------|--------------------------------|-----------------------------|-----------|--------------------|
| Bandar Abbas (south)   | 26.8                           | 211                         | Arid      | 0.09               |
| Jask (eastern south)   | 27.4                           | 132                         | Arid      | 0.04               |
| Lengeh (west)          | 27.8                           | 159                         | Arid      | 0.08               |
| Minab (east)           | 28.6                           | 201                         | Arid      | 0.10               |
| Ahmadi (eastern north) | 19.2                           | 310                         | Semi-arid | 0.20               |

P/ET<sup>o</sup>: ratio of mean annual precipitation to mean annual reference crop evapotranspiration.

southwest. In this zone, the Late Precambrian–Middle Triassic, Carbonatic sequence of Jurassic–Cretaceous, and Synorogenic Cenozoic sequence of marine or nonmarine nature are the lithostratigraphic units recognized. (ii) Makran zone, which extends from east to southeast. The basement of this zone, contrary to that of Zagros, is of oceanic and flysch-type sediments of Late Cretaceous–Pliocene. (iii) Central Iran is situated in the northern extremity of this province. Central Iran subzone rock sequences are deformed and metamorphosed (Aghanabati, 2004).

During the major part of the Phanerozoic era, the Persian Gulf Basin was located on an ancient passive margin of Gondwana, which opened toward the Paleotethys Ocean in the Paleozoic era and toward the Neotethys in the Mesozoic era (Ziegler, 2001). The stable subsidence and the unique landscape-climatic conditions favored the accumulation of a very thick sedimentary lens of carbonate rocks and evaporites. Carbonate rocks with excellent hydrocarbons reservoir properties are widespread. Basement rocks in the majority of the region are overlain by different age sandstones, which overlie salts and limestones in some areas. The age of these salts known as the Hormuz Salt is explained by different authors as Late Precambrian or Early Cambrian (Sharland et al., 2001). For the eastern zone, the Makran mountain chain, the most important geological limitation is the erosion of its geological structures. The oldest rocks in this region belong to the Cretaceous. After that, the Ophiolite complex sediments of Eocene–Oligocene can be seen. These sediments have Ophiolite facies and consist of sandstone, shale, and clay (Konyuhov and Maleki, 2006).

## Laboratory Analyses

### Physicochemical Analysis

Soil samples from different soil horizons ( $n = 41$ ), depicting most representative soil types, terrain units, and climatic zones (Fig. 1B), as described in Soil Survey Staff (1999), and corresponding sedimentary parent rocks ( $n = 16$ ) were selected for the study. Air-dried samples were crushed to pass a 2-mm sieve for analysis. Soil properties and analyzed constituents include particle size distribution using the hydrometer method described by Bouyoucos (1962); organic matter (OM) content by wet combustion (Nelson and Sommers, 1996); cation exchange capacity (CEC) by 1*N* sodium acetate at pH 8.2 (Chapman, 1965); percentage of gypsum by using acetone for precipitation (Soil Conservation Service, 1972) and calcium carbonate equivalent (CCE) by titration (Salinity Laboratory Staff, 1954); and soluble  $\text{Ca}^{2+}$ ,  $\text{Mg}^{2+}$ ,  $\text{Na}^+$ , electrical conductivity (EC), and pH in saturated paste extracted according to Bashour and Sayegh (2007). Soils were classified as Aridisols, Entisols, and Inceptisols, with calcic, gypsic, salic, cambic, and argillic horizons and ochric epipedon (Soil Survey Staff, 2010).

### Mineralogical Analyses

A representative fraction of the samples was used for the mineralogical study of the bulk soil and rock samples, and another

was used for the extraction and study of the clay, silt, and sand in fine earth (<2 mm) fraction. For mineralogical analysis of extracted particles, soil samples were oven dried. After removal of gypsum by repeated dissolution in distilled water, carbonates, OM, and Fe oxides were also removed using 1*N* sodium acetate buffered at pH 5, 30%  $\text{H}_2\text{O}_2$ , and citrate-bicarbonate-dithionate, respectively (Jackson, 1979). Eight samples from Cretaceous, Paleocene, Miocene, and Pleistocene sediments as the representative soil parent materials were also taken and prepared for mineralogical studies. Next, selected soil particle size fractions were isolated for mineralogical analyses. Clay fractions of the treated samples were separated by centrifugation (Jackson, 1979), and oriented slides were prepared for the treatments: heating to 550°C, solvation with glycerol, solvation with dimethyl sulfoxide (Moore and Reynolds, 1997). The prepared samples were scanned using an X-ray diffractometer (Philips PW-1710  $\text{CuK}\alpha$  radiation), and X-ray diffraction (XRD) patterns were obtained for air-dried ground rock, soil, silt, and sand samples according to Azaroff and Buerger (1958). Interpretation of XRD diffractograms and quantitative mineralogy was executed by Xpovder program, version 2004.04 ([www.xpowder.com](http://www.xpowder.com)).

For transmission electron microscope (TEM) studies, homogeneous diluted clay suspensions on pure ethyl alcohol were dried on Cu/Au grids and studied by a Philips CM-20 microscope operated at 200 kV (fitted with an ultrathin window, solid-state Si(Li) detector for energy-dispersive X-ray analysis). An attached energy-dispersive X-ray spectroscopy (EDX) detector was used for microanalyses and a CCD camera for image capturing. Chemical composition of palygorskite was carried out using analytical electron microscopy. The results of more than 1,000 analytical electron microscopy analyses and structural formulae were obtained for the samples studied. The structural formulae for all analyzed particles were calculated on the basis of  $\text{O}_{20}(\text{OH})_2$ .

For thin-section studies, selected rock slabs were vacuum impregnated with epoxy resin (for porous and soft samples) to preserve textural integrity during preparation of more than 20 thin sections. Thin sections were studied under normal polarized and also with circular polarized light with petrographic microscope. Micromorphological descriptions are based on criteria and terminologies of Bullock et al. (1985) and Stoops (2003).

## RESULTS AND DISCUSSION

### Soil Properties and Classification

The location of soil profiles and their classification are given in Table 2. Physiographically, soils were located on lowlands, floodplains, flood basins, plateaus, alluvial plains, piedmont plains, and fans. The soils on fans and piedmonts are shallow to medium deep, excessively drained, sandy-skeletal to loamy skeletal, highly calcareous, and severely eroded, with low organic carbon and

**TABLE 2.** Profiles, Classification, and General Field Aspects of the Studied Soils

| Profile Position | Profile No. | Physiography           | Soil Classification       | Soil Color   |
|------------------|-------------|------------------------|---------------------------|--------------|
| North            | 25          | Floodplain             | Typic Haplocambids        | 10 YR 5/3    |
| North            | 26          | Floodplain             | Typic Calcustepts         | 10 YR 4/4    |
| North            | 20          | Alluvial plain         | Typic Haplocambids        | 10 YR 6/4    |
| North            | 27          | Piedmont               | Lithic Ustic haplocalcids | 10 YR 6/3    |
| North            | 28          | Plateau                | Ustic Calcicgirds         | 10 YR 3.5/4  |
| Eastern north    | 13          | Plateau                | Fluventic Haplocambids    | 10 YR 6/3    |
| Eastern north    | 14          | Piedmont               | Typic Torriorthents       | 10 YR 7/5    |
| Eastern north    | 24          | Alluvial Plain         | Typic Haplocambids        | 10 YR 6.5/4  |
| Eastern north    | 9           | Plateau                | Torreric Calcustepts      | 10 YR 5/2    |
| South            | 19          | Alluvial Plain         | Typic Haplogypsis         | 7.5 YR 6/3   |
| South            | 18          | Piedmont               | Sodic Haplogypsis         | 7.5 YR 6/4   |
| South            | 17          | Colluvial fan          | Typic Torrifluvents       | 10 YR 7/4    |
| South            | 11          | Floodplain             | Typic Haplocambids        | 10 YR 6.5/5  |
| South            | 10          | Floodplain             | Typic Haplocalcids        | 10 YR 5/4    |
| East             | 1           | Lowland                | Typic Endoaquents         | 2.5 Y 5/2    |
| East             | 2           | Flood basin            | Gypsic Haplosalids        | 7.5 YR 7/5   |
| East             | 5           | Floodplain             | Typic Haplosalids         | 7.5YR 7/4    |
| East             | 6           | Flood basin            | Ustic Haplocambids        | 10 YR 5/4    |
| East             | 7           | Piedmont               | Ustic Haplocalcids        | 10 YR 6/3.5  |
| East             | 8           | Piedmont               | Ustic Calcigypsis         | 7.5 YR 7/5   |
| West             | 23          | Floodplain             | Typic Aquisalids          | 5 YR 5/2     |
| West             | 22          | Floodplain             | Typic Haplogypsis         | 2.5 YR 7.5/2 |
| West             | 4           | Floodplain             | Typic Haplogypsis         | 7.5 YR 6/3   |
| West             | 3           | Alluvial plain         | Gypsic Haplosalids        | 10 YR 6/6    |
| West             | 12          | Floodplain             | Typic Haplocambids        | 10 YR 6/5    |
| West             | 16          | Piedmont               | Leptic Haplogypsis        | 7.5 YR 6/4   |
| West             | 15          | Alluvial-colluvial fan | Typic Torriorthents       | 7.5 YR 7/5   |

available water capacity. In general, soils with continuous sedimentation, medium texture, and high salinity are settled in floodplains and lowlands. Moderate subangular blocky to angular blocky and massive structures are common; however, some horizons have platy, prismatic, or columnar structures. Prismatic and columnar structures are a common structural type in sodic soils (Soil Survey Staff, 2010).

In the modern US Soil Taxonomy (Soil Survey Staff, 2010), the presence of less developed diagnostic horizons (cambic, ochric) and the absence of well-developed horizons (temporary argillic, oxic, etc.) in soils are considered to be the result of the limiting influence of climatic (Aridisols), lithological, or chronological factors (Inceptisols, Entisols). Within these orders, relatively more developed soils occur at the levels of suborders and soil great groups, reflecting the total results of the “opposition” of this or that soil-forming factor to the principal factor limiting soil development. However, the nature of the bedrock (calcareous and gypsiferous) appeared to determine the additional presence of certain characteristic taxonomic units that may suggest the presence of locally adapted groups.

The soils have Munsell colors ranging from 10YR to 2.5YR (Table 2). Soil matrix colors are very different between each location, with the heavier soils having darker colors. The majority of soils are classified as Aridisols and Entisols with regard to the aridic soil moisture regime common in the study area (Table 2). The Inceptisols with calcic horizons were only reported in north and northeastern regions of Hormozgan Province. Gypsiferous soils increase as we move toward the western section and, in

comparison, soils with a calcic horizon increase toward the northern sector.

Calcic, gypsic, salt, and textural features are the main morphological features in the studied pedons. Calcic features were observed in the field as powdery pockets, nodules, or concretions in B<sub>tk</sub> or B<sub>k</sub> horizons that have formed in plateaus, piedmonts, and floodplains. Pendants, powdery pockets, and crystals of gypsum are the main gypsum features, whereas laminar to puffy salt crusts and salt crystals are among salt features in the field. Clay skins were noted in the field on ped surfaces attributed to argillic horizons that are associated with calcic horizons. Occurrence of an argillic horizon in calcareous parent material of an arid climate could point to a paleoprocess in stable geomorphologic conditions (as plateau) (Khormali et al., 2003).

### Soil Characteristics and Particle Size Distribution of Soils

The particle size fractions, pH, EC, sodium adsorption ratio (SAR), CEC, and percentages of CCE, gypsum, and OM of the 41 soil samples are given in Table 3. The pH of soils varied from 7.1 to 8.7, with an average of 7.8. Because the soils developed on highly carbonatic and gypsic bedrocks, the pH values are within the normal range for these types of soils. The high values could also be a result of impression by salt formations. Also, irrigation water contains significant quantities of calcium, magnesium, or sodium in certain regions. On the other hand, these soils generally occur in areas where precipitation is too low to leach the

TABLE 3. Selected Physicochemical Properties of the Studied Soils ( $n = 41$ )

| Pedon No. | Hor. | Sand | Silt | Clay | pH   | EC, dSm <sup>-1</sup> | SAR                    | CEC, cmol <sub>c</sub> kg <sup>-1</sup> | CCE   | Gypsum | OM    |
|-----------|------|------|------|------|------|-----------------------|------------------------|---|-------|--------|-------|
|           |      | %    |      |      |      |                       |                        |   | %     |        |       |
| 1         | Cgz2 | 12.5 | 82   | 5.5  | 8.31 | 44.5                  | 14.36                  | 8.57                                    | 32.5  | 0.1    | 0.66  |
|           |      |      |      |      |      |                       | Typic Endoaquents      |   |       |        |       |
| 2         | Byz  | 40   | 50   | 10   | 7.21 | 45                    | 8.42                   | 12.2                                    | 39.2  | 11.4   | 0.5   |
| 2         | C    | 62   | 30   | 8    | 7.7  | 45                    | 6.5                    | 11.9                                    | 48    | 3.3    | 0.45  |
|           |      |      |      |      |      |                       | Gypsic Haplosalids     |   |       |        |       |
| 3         | Byz  | 34.5 | 61.5 | 4    | 7.24 | 24                    | 3.07                   | 6.09                                    | 33.57 | 9.7    | 0.86  |
|           |      |      |      |      |      |                       | Typic Haplogypsis      |   |       |        |       |
| 4         | By1  | 30   | 66   | 4    | 7.68 | 8.5                   | 1.51                   | 12.2                                    | 38.6  | 10.11  | 1.1   |
|           |      |      |      |      |      |                       | Typic Haplosalids      |   |       |        |       |
| 5         | Cz   | 36   | 60   | 4    | 8.55 | 27.5                  | 8.55                   | 5                                       | 21.1  | 10     | 1.1   |
|           |      |      |      |      |      |                       | Ustic Haplocambids     |   |       |        |       |
| 6         | Bw   | 58   | 40   | 2    | 8.04 | 0.45                  | 0.12                   | 5.43                                    | 37    | 0.2    | 0.6   |
|           |      |      |      |      |      |                       | Ustic Haplocalcids     |   |       |        |       |
| 7         | Bk   | 20   | 75   | 5    | 7.62 | 2.5                   | 1.35                   | 4.7                                     | 30.23 | 0.17   | 1.12  |
|           |      |      |      |      |      |                       | Ustic Calcigypsis      |   |       |        |       |
| 8         | By   | 63   | 33   | 4    | 7.83 | 3.3                   | 0.05                   | 3.95                                    | 77.4  | 10.3   | 0.86  |
|           |      |      |      |      |      |                       | Torreric Calcustepts   |   |       |        |       |
| 9         | A    | 59   | 39   | 4    | 7.76 | 6.4                   | 1.0                    | 10                                      | 39.2  | 2.15   | 0.75  |
| 9         | Bk1  | 52   | 44   | 4    | 7.89 | 5.0                   | 0.03                   | 11                                      | 56.3  | 3.4    | 0.59  |
| 9         | C    | 47   | 49   | 4    | 7.94 | 5.1                   | 0.03                   | 9                                       | 61.0  | 3.9    | 0.55  |
|           |      |      |      |      |      |                       | Typic Haplocalcids     |   |       |        |       |
| 10        | Bk1  | 53   | 37   | 10   | 7.76 | 3.7                   | 0.00                   | 4.5                                     | 42.5  | 7.03   | 0.22  |
| 10        | Bk2  | 43   | 52   | 5    | 7.66 | 3.1                   | 0.02                   | 5                                       | 40.6  | 8      | 0.35  |
|           |      |      |      |      |      |                       | Typic Haplocambids     |   |       |        |       |
| 11        | A    | 69   | 28   | 3    | 7.63 | 18.5                  | 1.0                    | 6                                       | 39.6  | 1.8    | 0.5   |
| 11        | C    | 86   | 12   | 2    | 7.97 | 2.35                  | 0.02                   | 9.1                                     | 29.1  | 0.1    | 0.0.7 |
|           |      |      |      |      |      |                       | Typic Haplocambids     |   |       |        |       |
| 12        | Bw   | 53   | 42   | 5    | 7.97 | 8.5                   | 0.2                    | 8.44                                    | 49.2  | 0.38   | 0.63  |
|           |      |      |      |      |      |                       | Fluventic Haplocambids |   |       |        |       |
| 13        | Bw   | 59   | 38   | 3    | 8.6  | 3.45                  | 0.06                   | 8.6                                     | 35.7  | 0.35   | 0.57  |
|           |      |      |      |      |      |                       | Typic Torriorthents    |   |       |        |       |
| 14        | Ap   | 69   | 29.5 | 1.5  | 7.66 | 10.5                  | 0.03                   | 9.15                                    | 35.13 | 0.75   | 0.45  |
| 14        | C    | 70   | 24.5 | 5.5  | 8.27 | 13                    | 0.9                    | 16.32                                   | 28.44 | 0.1    | 0.43  |
|           |      |      |      |      |      |                       | Typic Torriorthents    |   |       |        |       |
| 15        | Ap   | 80   | 18   | 2    | 7.71 | 7                     | 0.01                   | 7.23                                    | 29.22 | 0.37   | 0.30  |
| 15        | C    | 83   | 15   | 2    | 8.41 | 7.75                  | 0.22                   | 5.69                                    | 25.88 | 0.6    | 0.20  |
| 15        | C    | 83   | 15   | 2    | 8.41 | 7.75                  | 0.22                   | 5.69                                    | 25.88 | 0.6    | 0.20  |
|           |      |      |      |      |      |                       | Leptic Haplogypsis     |   |       |        |       |
| 16        | Ap   | 60   | 39   | 1    | 7.74 | 2.85                  | 0.1                    | 5.82                                    | 73.04 | 14     | 0.74  |
| 16        | By   | 90   | 8    | 2    | 8.08 | 5.6                   | 0.02                   | 5.96                                    | 53.6  | 26.9   | 0.37  |
|           |      |      |      |      |      |                       | Typic Torrifluvents    |   |       |        |       |
| 17        | C2   | 86   | 12   | 2    | 8.43 | 0.87                  | 0.03                   | 2.15                                    | 81.53 | 0.45   | 0.22  |
|           |      |      |      |      |      |                       | Sodic Haplogypsis      |   |       |        |       |
| 18        | By3  | 52.5 | 42   | 5.5  | 7.6  | 19.5                  | 2.39                   | 4.67                                    | 73    | 25.8   | 0.74  |
|           |      |      |      |      |      |                       | Typic Haplogypsis      |   |       |        |       |
| 19        | By   | 56   | 46   | 8    | 7.56 | 2.3                   | 0.10                   | 5.56                                    | 65.5  | 19.4   | 0.7   |
|           |      |      |      |      |      |                       | Typic Haplocambids     |   |       |        |       |
| 20        | A    | 64.5 | 32.5 | 3    | 7.88 | 0.54                  | 0.04                   | 8.92                                    | 61.45 | 0.1    | 0.78  |
| 20        | Bw2  | 64   | 32   | 4    | 8.47 | 1.3                   | 0.56                   | 4.18                                    | 79.1  | 0.17   | 0.78  |
|           |      |      |      |      |      |                       | Gypsic Haplosalids     |   |       |        |       |
| 21        | Byz2 | 75   | 8    | 17   | 8.3  | 39                    | 8.59                   | 5                                       | 67    | 18     | 0.64  |

Continued next page

TABLE 3. (Continued)

| Pedon No.                 | Hor. | Sand | Silt | Clay | pH   | EC, dSm <sup>-1</sup> | SAR  | CEC, cmol <sub>c</sub> kg <sup>-1</sup> | CCE   | Gypsum | OM   |
|---------------------------|------|------|------|------|------|-----------------------|------|---|-------|--------|------|
|                           |      | %    |      |      |      |                       |      |   | %     |        |      |
| Typic Haplogypsis         |      |      |      |      |      |                       |      |   |       |        |      |
| 22                        | A    | 34   | 62   | 4    | 7.87 | 0.022                 | 0.10 | 4.67                                    | 61.5  | 6      | 1.8  |
| 22                        | By   | 54   | 43   | 3    | 7.49 | 2.85                  | 0.13 | 6.79                                    | 47.5  | 28     | 0.9  |
| Typic Aquisalids          |      |      |      |      |      |                       |      |   |       |        |      |
| 23                        | Cg1  | 64   | 32   | 4    | 7.8  | 1.5                   | 0.82 | 9.57                                    | 55.42 | 1.6    | 0.4  |
| Typic Haplocambids        |      |      |      |      |      |                       |      |   |       |        |      |
| 24                        | Bw   | 56   | 40   | 4    | 7.63 | 5.85                  | 0.32 | 6.3                                     | 68.4  | 21.5   | 0.57 |
| 24                        | C    | 43   | 46   | 11   | 7.58 | 4.3                   | 0.89 | 5                                       | 66.2  | 21.5   | 0.76 |
| Typic Haplocambids        |      |      |      |      |      |                       |      |   |       |        |      |
| 25                        | Ap   | 32.5 | 60   | 7.5  | 7.45 | 18                    | 0.1  | 11                                      | 56.5  | 5      | 1    |
| Typic Calcustepts         |      |      |      |      |      |                       |      |   |       |        |      |
| 26                        | Bw   | 62   | 28   | 10   | 7.74 | 0.35                  | 0.97 | 7.08                                    | 38.5  | 0.08   | 0.57 |
| Lithic Ustic Haplocalcids |      |      |      |      |      |                       |      |   |       |        |      |
| 27                        | Bk   | 35   | 56   | 9    | 7.63 | 2.6                   | 0.78 | 11.37                                   | 51.41 | 0.14   | 0.66 |
| Ustic Calcargids          |      |      |      |      |      |                       |      |   |       |        |      |
| 28                        | A    | 18   | 70   | 12   | 7.08 | 3.8                   | 0.63 | 19                                      | 35    | 0.11   | 1.6  |
| 28                        | Bt   | 18   | 64   | 18   | 7.84 | 0.18                  | 0.78 | 10.25                                   | 41.4  | 0.32   | 1.2  |
| 28                        | Bk   | 24   | 62   | 14   | 7.9  | 0.47                  | 1.10 | 13                                      | 41.2  | 0.4    | 1.3  |

minerals from the soils. Soils with naturally high sodium levels or those that have received large quantities of sodium bicarbonate through irrigation may have pH levels above 8.3 as 8.6. Theoretically, if sodium is not a factor, even if large quantities of calcium or magnesium carbonate are applied, the soil pH will not exceed 8.2 to 8.3 (Snyder, 2007). In the natural system, the soil pH is affected by the minerals, climates, and weathering. Management of soils often alters the natural pH to acidic levels by the use of acid-forming fertilizers. Also, removal of bases (potassium, calcium, and magnesium) from the soil can occur under continuous irrigation management (Balba, 1995; NRCS, 1998).

The EC and SAR values of the soils are generally in a broad range, 0.02 to 45 dS m<sup>-1</sup> and 0.01 to 14.36, respectively. The high values of EC in pedons 1, 2, 3, 4, and 5 are related to saline groundwater and parent material; in pedons 9, 11, 12, 14, and

15, to saline irrigation water; and in pedons 18, 21, and 25, to parent materials containing saline composition and uplifting of salt diapirs (Hormoz Formation). Some soils contain significantly more sodium compared with calcium or magnesium, which may be a risk of sodium buildup on soils. In the studied soils, salinity and alkalinity occur together. As shown in Fig. 2, there was a significant correlation between calculated SAR and EC of soil samples. Thus, soils are classified into nonsaline and nonsodic (EC, <6; SAR, <8), saline (EC, >6; SAR, <8), and saline-sodic soils (EC, >6; SAR, >8) (Salinity Laboratory Staff, 1954).

Soils exhibited CEC between 2.15 and 18 cmol<sub>c</sub>kg<sup>-1</sup>. The CEC of the soils depends on the amount and type of clay minerals, OM, and gypsum content. Gypsum does not have electric charge; therefore, with increasing gypsum in the soils, CEC decreases. Low values of CEC in the soils are attributed to low OM, low clay

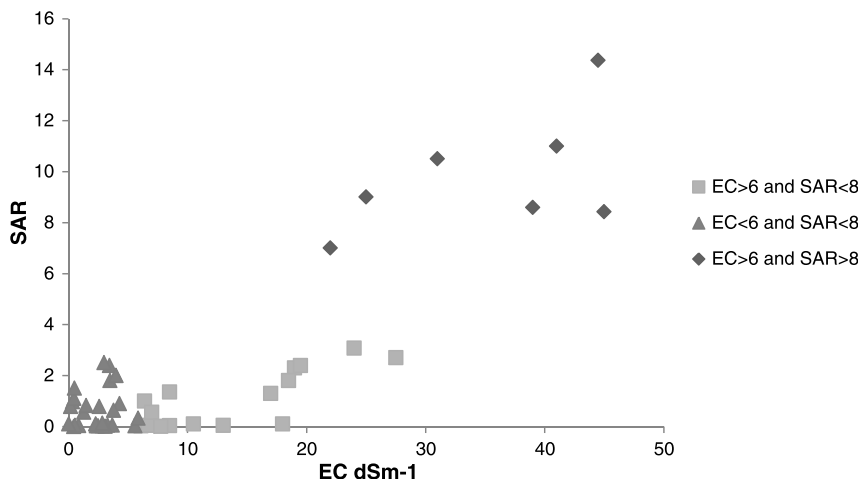


FIG. 2. Correlation between SAR and EC data in study soil samples ( $R^2 = 0.74$ ,  $P < 0.05$ ).

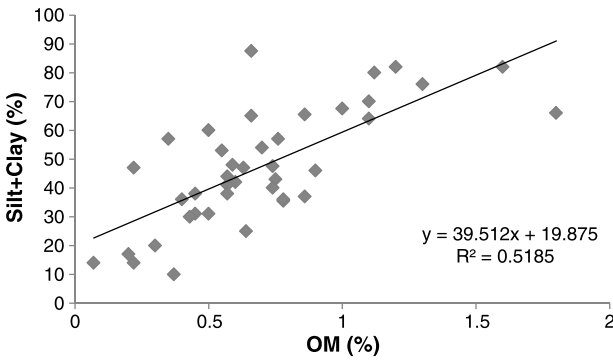


FIG. 3. Correlation between silt + clay and OM in study soil samples.

minerals with low CEC, and presence of gypsum in notable amounts in most soil samples. Organic matter varies from 0.07 to 1.8% and was positively correlated with clay + silt contents (Fig. 3), which is in accordance with the findings of Amelung et al. (1998) and Sultan (2006).

Soil samples are classified as sandy loam, silt loam, loamy sand, silt, sand, loam, and silty clay loam, with a low clay size fraction (Fig. 4). The average value for clay, silt, and sand size fractions were 5.8, 41.1, and 53.6%, respectively. Therefore, sand and silt are the most abundant fractions. Gypsum and CCE percentage varied from less than 1 to 29.0% and 21.1 to 81.5%, respectively. The soils in Hormozgan Province are dominated by calcium and magnesium carbonate, thus are classified as calcareous and gypsiferous soils.

## Mineralogical Characterization

### Soil Mineralogy

The soils contained calcite (29%), dolomite (19.5%), alkali-feldspars (19%), mica (muscovite or biotite) (15.6%), quartz (14.3%), gypsum (13%), 14 Å-phyllsilicates (11.6%), plagioclase (6.6%), and halite (<1%) minerals (Fig. 5). Most of the soils contain calcite and gypsum in the fine fraction (<2 mm). Dolomite was found in soils developed in north, northeastern, and west districts, with the highest amounts in samples from western regions with evaporate formations. The presence of dolomite in Zagros zone sediments compared with Makran zone could be related to precipitation of gypsum after evaporation of seawater. After precipitation of gypsum, the brine is both enriched in Mg, relative to Ca, and has a higher density. The brine then enters the groundwater system and moves downward into buried limestones. The Mg-rich brine then reacts with the calcite in the limestone to produce dolomite. Higher quantities of plagioclase were detected in samples that were under cultivation and relatively developed with low salinity (i.e., pedons 13, 14, 15, 27, and 28). Generally, the second and third most abundant mineral groups were alkali-feldspar and mica, respectively. Halite mineral ranged from 5 to 10% in salic soil horizons.

The rates of weathering and alteration, and thus the rates of soil (or sediment) accumulation and maturation, are governed by climate (temperature, the availability of H<sub>2</sub>O, biotic factors). The formation of a soil is also dependent on the bedrock type in the area, physical factors (such as rock porosity and texture), and mineralogical factors (solubility). The soils are also susceptible to many anthropogenic (human-induced) effects. The landscape distribution of calcite and gypsum minerals as well is often strongly

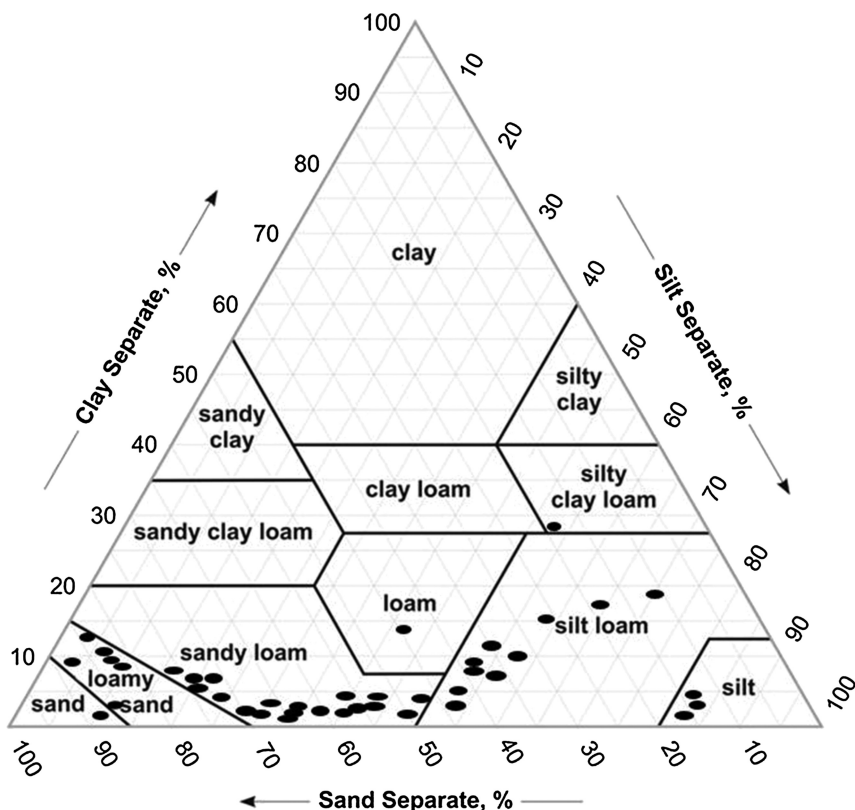


FIG. 4. Textural classification of the studied soils.

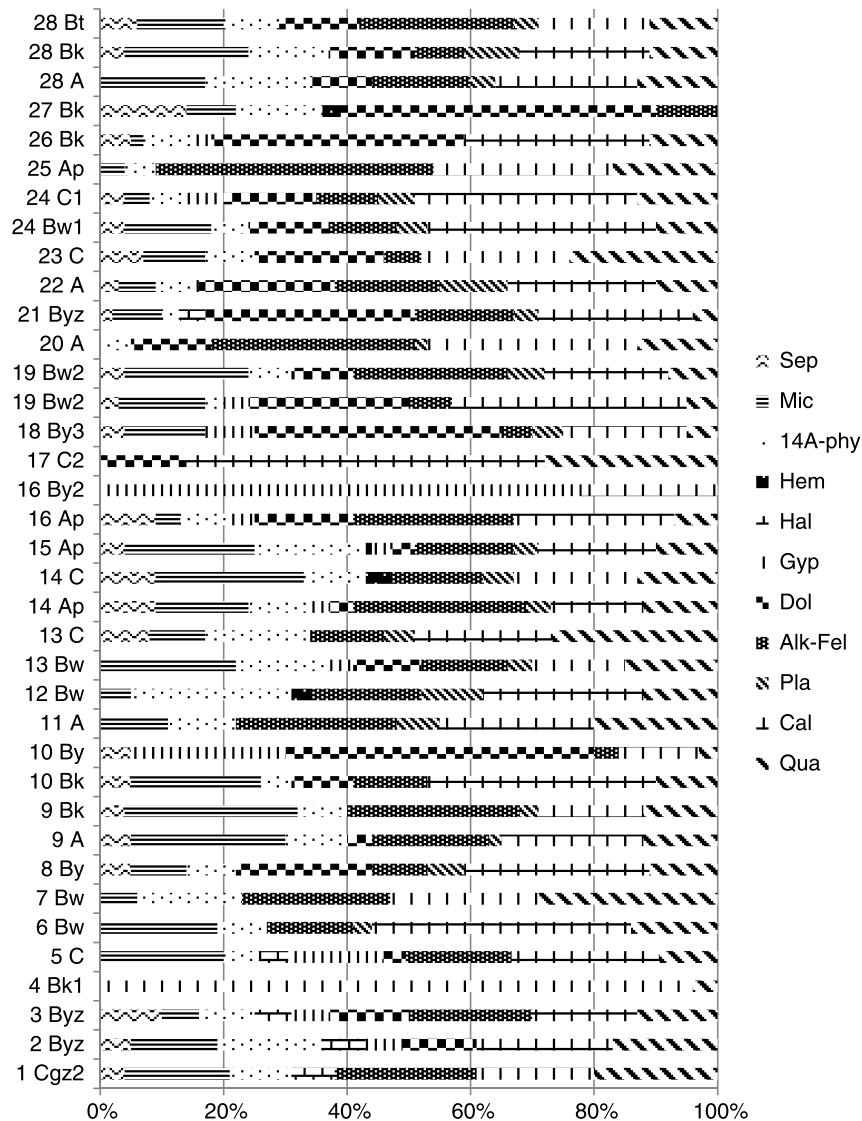


FIG. 5. Relative distribution of soil minerals in different soil horizons.

related to patterns of eolian dust deposition and water infiltration and leaching (Graham and O'Geen, 2010). The presence of secondary calcite and gypsum is consistent with the fact that the soil receives low precipitation. Minerals such as carbonates, gypsum, and soluble salts, common in Aridisols, may also be, at least in part, wind transported; however, their persistence in the soil is caused by limited leaching.

González et al. (2007) concluded that calcite, biotite, and most of all chlorite are the main minerals in the marls. The soils developed on them, with gypsum predominant in the gypsiferous marl in a diapiric formation north of Spain. The semiarid conditions in the study zone, together with the processes of extrusion and hydrothermal activity affecting the formation of the diapir, should be responsible for the genesis of the minerals.

Sepiolite is rarely detected in soils by XRD mainly because of its dissolution under NaOAC treatment at pH 4.5 for the removal of carbonate (Abtahi, 1977, 1985). Under acidic conditions, sepiolite is destroyed more rapidly than palygorskite because of its magnesian composition and the larger size of its structural

microchannels (Myriam et al., 1998). Moreover, sepiolite has a lower stability in the presence of high Al in soil, which provides favorable conditions for the formation of palygorskite rather than sepiolite. In soils, without acid treatment, sepiolite was identified as common clay mineral. This conclusion was based on the presence of strong peaks at 11.8 to 12.4 Å and moderate peaks at 4.1 to 4.5 Å. The high levels of evaporate minerals (salts, gypsum, and carbonates) in surface and upper horizons of soils as well as high values of feldspars show a low weathering rate and limited precipitation. Thus, it can be concluded that the soils have commonalities with parent materials. Limited quantities of the elements have been removed from the soils.

### Sand and Silt Mineralogy

Sand and silt size gradings conceal important geological processes because there are geological controls on both the quartz sand and silt fractions. The sand fraction was dominated by quartz, alkali-feldspar, 14 Å-phyllsilicates, plagioclase, mica, dolomite,



gypsum, and rare amounts of calcite (Fig. 6). Dolomite was part of the sand fraction in pedons 13, 16, and 20 sampled from the western areas. Nearly all the soils studied contain quartz as part of the sand fraction and the major part of 14 Å-phyllsilicates detected in northern and east-northern districts.

However, the silt fraction of the soils was characterized by a rather high amount of quartz, alkali-feldspars, mica, 14 Å-phyllsilicates, plagioclase, dolomite, and a very rare amount of calcite (Fig. 7). Quartz, alkali-feldspar, 14 Å-phyllsilicates, dolomite, and mica were present in silt fractions of all analyzed samples. The minerals of the silt fraction compared with minerals of the sand fraction had the same distribution and frequency. The only difference was the detection of gypsum minerals in the silt fraction that were not present in the sand fraction.

Parry et al. (2011) noted that the 2- to 63-µm fractions contributed up to 56 and 67% of bulk soil volume and BET surface area, respectively. Consideration of these observations and the mineralogy of this fraction suggest that the 2- to 63-µm fractions may be the most influential for the release of elements via mineral dissolution in the bulk soil. Silt fractions of the soils of the Wadi ar Rimah and side wadis, classified as Torrifluvents, are characterized by high quartz contents and the universal occurrence of kaolinite and feldspar, generally with plagioclase predominant; mica, chlorite, and hornblende also generally occur (Mashhady et al., 1980). The dunes are composed predominantly of well-sorted fine sands, consisting of unstrained quartz and carbonate grains, together with minor proportions of chert and feldspars (Abu-Zeid et al., 2001).

**Clay Mineralogy**

X-ray diffraction analysis of clay indicated that chlorite, palygorskite, illite, interstratified minerals (IM), and smectite are dominant clay minerals in the study area (Table 4). Accessory

minerals were quartz and feldspars. Traces of kaolinite with strong peaks at 11 to 11.5 Å on dimethyl treatment were found in pedons 12 and 19 from the southern and northeastern parts of the region (Fig. 8).

The occurrence of secondary clay minerals other than kaolinite suggests a climatic influence (Graham and O'Geen, 2010). Palygorskite, smectite, and vermiculite occur in soils of dry-hot environments, whereas hydroxy-interlayered minerals, gibbsite, and short-range ordered minerals are in soils of cold-moist environments. Broad relationships of soil clay mineralogy are often best correlated with mean annual precipitation rather than temperature (Birkeland, 1999). The amount of precipitation is important because percolating water leaches silica from soils and differentiates the type of clays that can form (Graham and O'Geen, 2010).

The limited leaching may favor the formation of Si-rich phyllosilicates, such as smectite, whereas kaolinite, if present, may be inherited or have been formed under wetter conditions than currently occurring. Occurrence of alkaline environments rich in Mg with an abundance of Si may also favor the synthesis of palygorskite and sepiolite (Singer, 1989). The high portion of palygorskite was probably caused by the soils classified as Gypsis to Salids. Soil pedons (6, 10, 16, 17, 20, 22, 23, 24, 26, and 27) developed on evaporate formations with gypsum, saline, and calcic materials and had elevated palygorskite content. The detailed mineralogy of palygorskite as a dominant and important mineral in arid study region is the focus of the next section. Sandy soils in Namaqualand contain palygorskite and sepiolite of pedogenic origin (Singer et al., 1995). The second dominant clay mineral was chlorite, which ranged from 15 to 60%. The higher contents were detected in soils developed in the northeastern and northern parts of the study area. Illite contents ranged between less than 2 and 35%. Mica, particularly biotite, and chlorite are commonly inherited clay minerals.

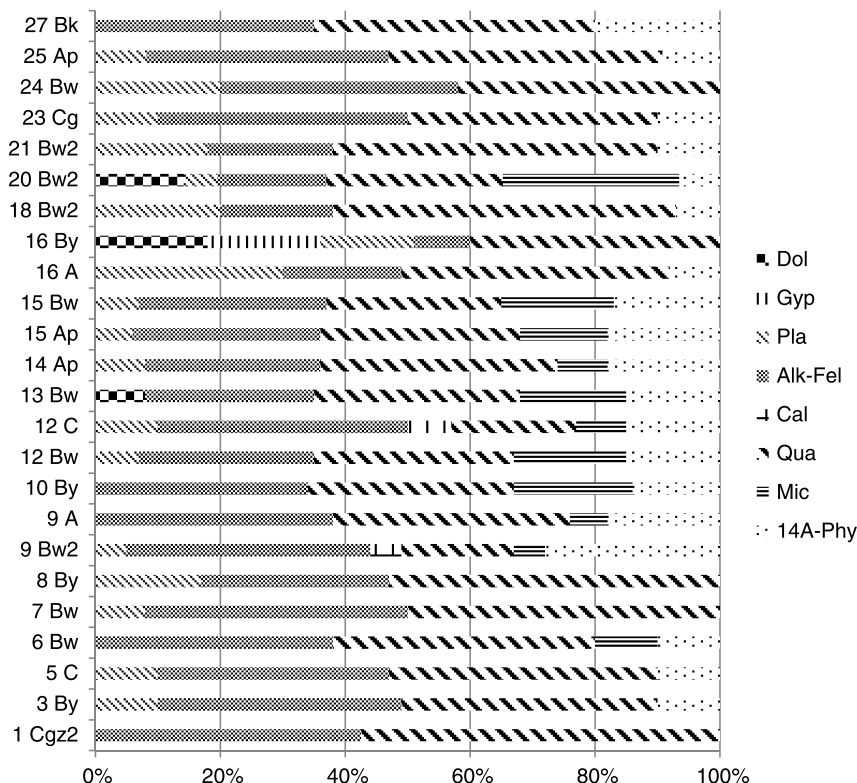


FIG. 6. Relative distribution of sand minerals in different soil horizons.

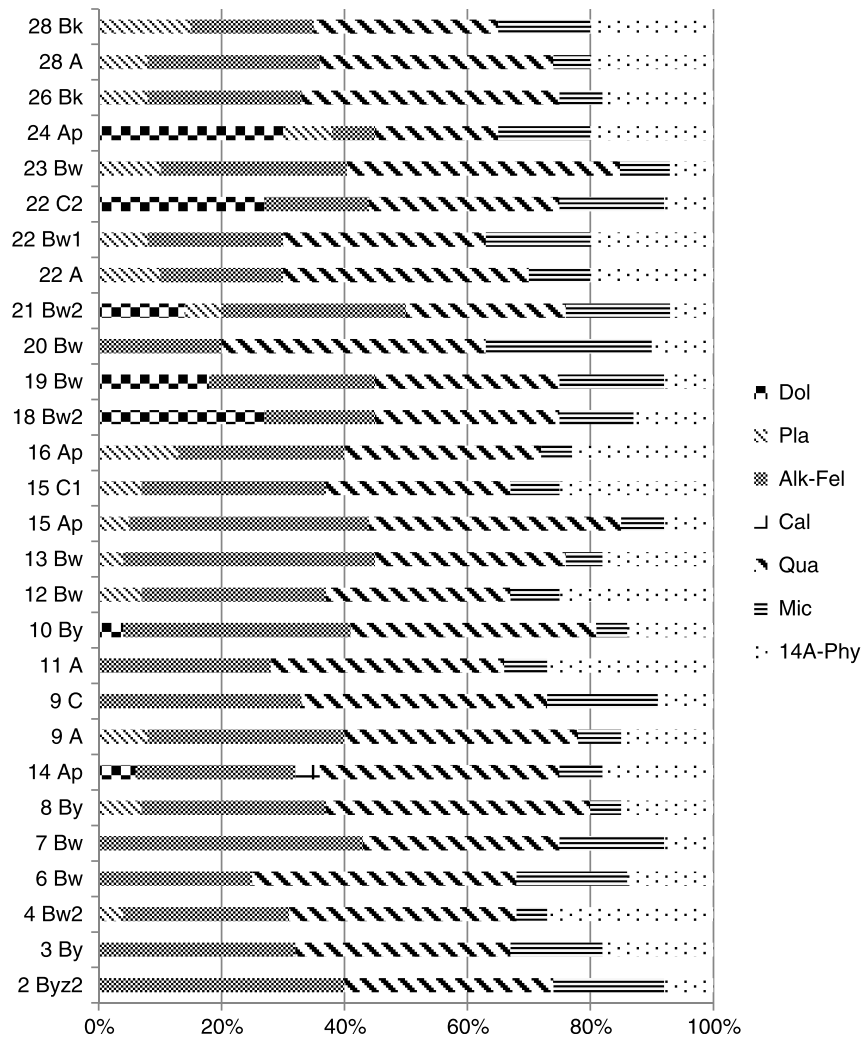


FIG. 7. Relative distribution of silt minerals in different soil horizons.

The high amounts of smectite and IM were detected in the northern and northeastern parts of the region. It can be concluded that the presence of relatively high amounts of smectite in saline soils is related to the precipitation from oversaturated solutions rich in Ca, Mg, and K under saline conditions. These elements are incorporated in the authigenically formed carbonates, smectites, and palygorskite in the soil profiles. The control of Ca occurs by precipitation of calcite and dolomite in nodules, whereas Mg and K are mainly involved in the neoformation of Mg-smectites.

The clay minerals are accompanied by palygorskite, illite, chlorite, smectite, mixed layer minerals, and, very rarely, kaolinite. Up to 75% of clay fractions are dominated by illite, chlorite, and palygorskite. The highest concentrations of the fibrous clay minerals commonly occur in the bottom horizons (Singer et al., 1995). Detrital input is the dominant factor responsible for the clay mineral distribution in sediments and soils. Palygorskite and associated smectite typically constitute terrigenous species.

Kaolinite was identified in the northeastern (profile 12) and southern parts of the region (Qeshm Island, profile 19). Quartz is also predominant in the clay fraction, with an average of 10%. In rare cases, quartz constitutes less than 2% of the clay fraction. Soil profiles that are dominated by kaolinite in the phyllosilicate mineral fraction are typically associated with environments

characterized by intense chemical weathering of soil materials in stable landscapes under warm humid conditions (Wilson, 1999). However, kaolinite may also be present in soil profiles as a result of inheritance from parent material (Southard and Miller, 1966). Therefore, the occurrence of kaolinite in the soil horizons and rocks is not surprising and indicates inheritance from the Paleocene time, indicating that warm and humid climatic conditions prevailed and persisted throughout the Late Pleistocene and Holocene in the region. Khademi and Arocena (2008) concluded that the occurrence of kaolinite in arid soils does not appear to be entirely caused by inheritance from parent materials. A portion of kaolinite in these soils could be neoformed from the transformation of fibrous clays in rhizospheres, a phenomenon not yet suggested by pedologists or soil mineralogists. Because rare amounts of kaolinite have been detected and also agricultural activities have been implemented only for hundreds of years in studied samples, kaolinitization of fibrous clays in the rhizosphere may not be a substantial process.

The major pathways for the formation of chlorite, illite, and kaolinite in the arid region of southern Iran with gypsiferous, calcareous, and saline soils are considered to be mainly of inherited origin from parent materials. Both inheritance and transformation from mica (illite) are concluded to be the main mechanisms for the

**TABLE 4.** Relative Mineral Components of Selected Clay Fraction as Determined by X-Ray Diffraction<sup>#</sup>

| Pedon | Horizon | Chl-Sme* | Ill-Sme* | Sme* | Chl* | Pal* | Ill* | Kao* |
|-------|---------|----------|----------|------|------|------|------|------|
| 1     | Cg1     | +        |          | +    | +++  | ++   | ++   | -    |
| 2     | Byz     | -        | -        | -    | ++++ | ++   | +++  | -    |
| 3     | By      | -        | +        | ++   | ++   | ++   | ++   | -    |
| 4     | Bw2     | -        | -        | -    | +++  | +++  | +++  | -    |
| 5     | C       | -        | -        | -    | +++  | +++  | +++  | -    |
| 6     | Bw      | -        | -        | -    | ++   | ++++ | +++  | -    |
| 7     | Bw      | -        | -        | ++   | +++  | ++   | ++   | -    |
| 8     | Bk      | +        | +        | -    | +++  | ++   | ++   | -    |
| 9     | A       | +        | -        | -    | +++  | +++  | ++   | -    |
| 9     | Bk1     | +        | ++       | -    | +++  | +++  | ++   | -    |
| 9     | C       | +        | +        | +    | +++  | +++  | ++   | -    |
| 10    | By      | -        | +        | -    | ++   | +++  | +++  | -    |
| 10    | Bk      | -        | -        | ++   | ++   | +++  | ++   | -    |
| 11    | A       | +        | +        | ++   | ++   | ++   | ++   | -    |
| 12    | Bw      | -        | -        | -    | ++++ | +++  | -    | ++   |
| 13    | Bw      | -        | -        | -    | ++++ | +++  | ++   | -    |
| 14    | A       | -        | -        | -    | +++  | +++  | ++   | -    |
| 14    | C       | +        | -        | ++   | +++  | +    | ++   | -    |
| 15    | Ap      | +        | +        | +    | +++  | ++   | ++   | -    |
| 16    | Ap      | -        | -        | -    | ++   | ++++ | +++  | -    |
| 16    | By      | -        | -        | +    | +++  | +++  | ++   | -    |
| 17    | C2      | -        | -        | -    | +++  | +++  | +    | -    |
| 18    | Bw2     | +        | +        | -    | +++  | ++   | +++  | -    |
| 19    | Bw2     | +        | -        | +++  | ++   | ++   | ++   | +    |
| 20    | A       | -        | -        | -    | +++  | +++  | +++  | -    |
| 21    | Bw2     | -        | -        | +++  | ++   | +++  | ++   | -    |
| 22    | Bw2     | +        | -        | +    | ++   | +++  | ++   | -    |
| 23    | C       | -        | -        | -    | +++  | ++++ | -    | -    |
| 24    | Bw      | ++       | +        | +    | ++   | +++  | ++   | -    |
| 25    | Ap      | +        | +        | +    | ++   | +++  | +++  | -    |
| 26    | Bk      | -        | -        | -    | +++  | +++  | +++  | -    |
| 27    | Bk1     | -        | +        | -    | +++  | +++  | +++  | -    |
| 28    | A       | +        | +        | -    | +++  | ++   | ++   | -    |
| 28    | Bt1     | ++       | ++       | +    | +++  | +    | +++  | -    |
| R1**  |         | -        | -        | -    | ++++ | -    | +++  | -    |
| R2    |         | -        | -        | -    | ++++ | -    | +++  | -    |
| R4    |         | -        | -        | ++   | ++   | ++   | ++   | -    |
| R5    |         | -        | -        | ++   | +++  | +    | ++   | +    |
| R6    |         | -        | -        | ++   | +++  | -    | ++   | -    |
| R7    |         | ++       | -        | +++  | ++   | ++   | ++   | -    |
| R8    |         | -        | -        | -    | ++++ | ++   | +    | -    |

\*Ill-sme: illite-smectite; chl-sme: chlorite-smectite; Sme: smectite; chl: chlorite; Pal: palygorskite; Ill: illite; Kao: kaolinite.

\*\*Rock samples.

<sup>#</sup>The abundance of components as estimated from XRD: ++++: >50%; +++: 25 to 50%; ++: 10 to 25%; +: 2 to 10%; -: not detected.

occurrence of smectite in the other studied pedons. Neof ormation of palygorskite was the major mechanism for the occurrence of this clay mineral in the studied soils, these results are in accordance with the findings of Azizi et al. (2011) in\* gypsiferous soils of southern Tehran (Iran).

### Palygorskite

The presence of fibrous minerals (palygorskite) in clay-sized particles was detected by TEM equipped with energy-dispersive

X-ray analysis as the most abundant clay minerals in the collected samples (Fig. 9). Palygorskite and sepiolite are fibrous clay minerals occurring in the form of sedimentary rock deposits, hydrothermal or metamorphic reaction products in altered magmatic rocks, or as components of the clay (<2 μm) fraction in soils from arid or semiarid areas. Palygorskite is a common clay mineral in soils and sediments of arid and semiarid areas and is unstable in areas with precipitation levels greater than 300 mm (Singer, 1989).

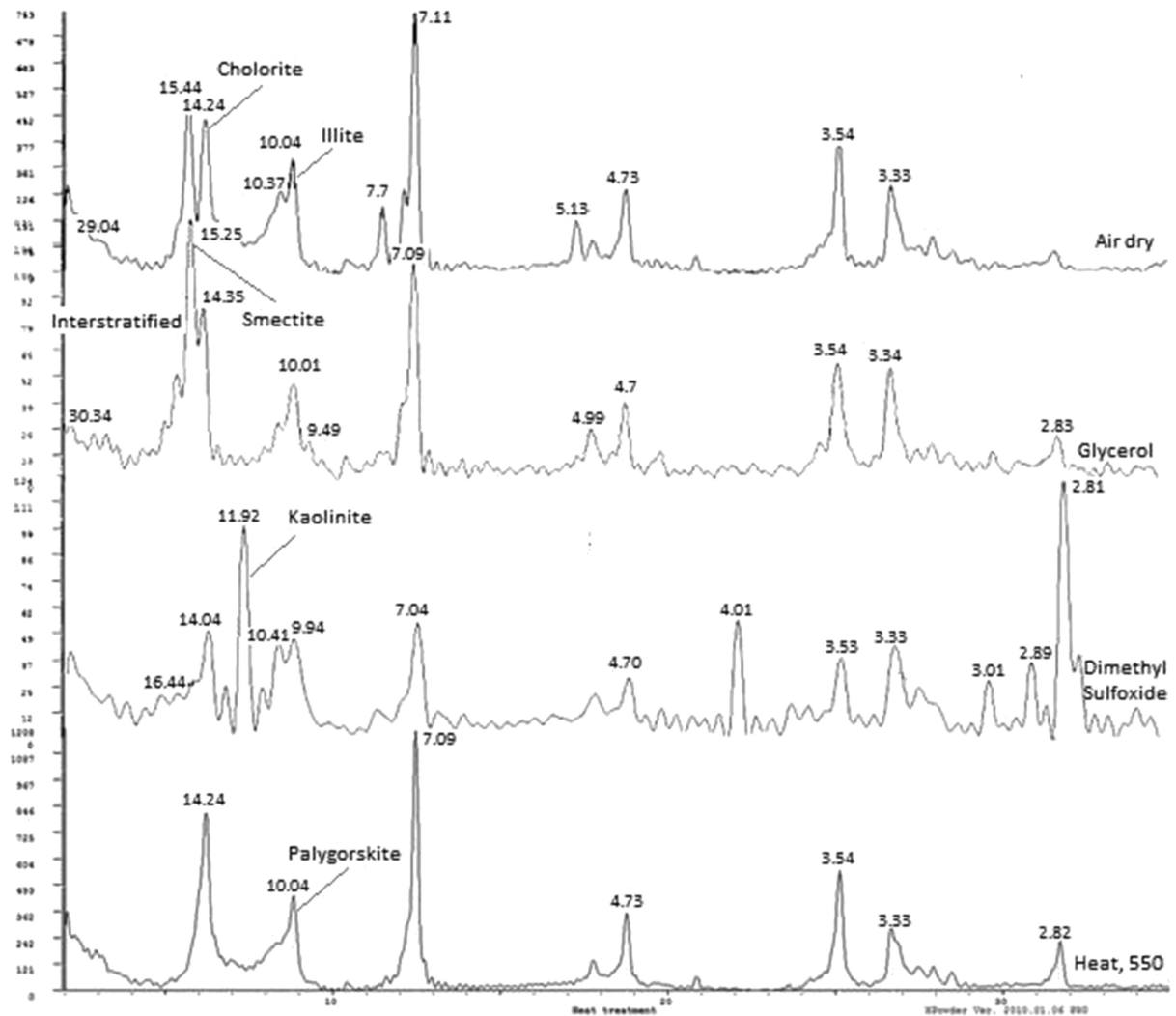


FIG. 8. X-ray diffractograms of the clay fraction (sample 19 Bw2).

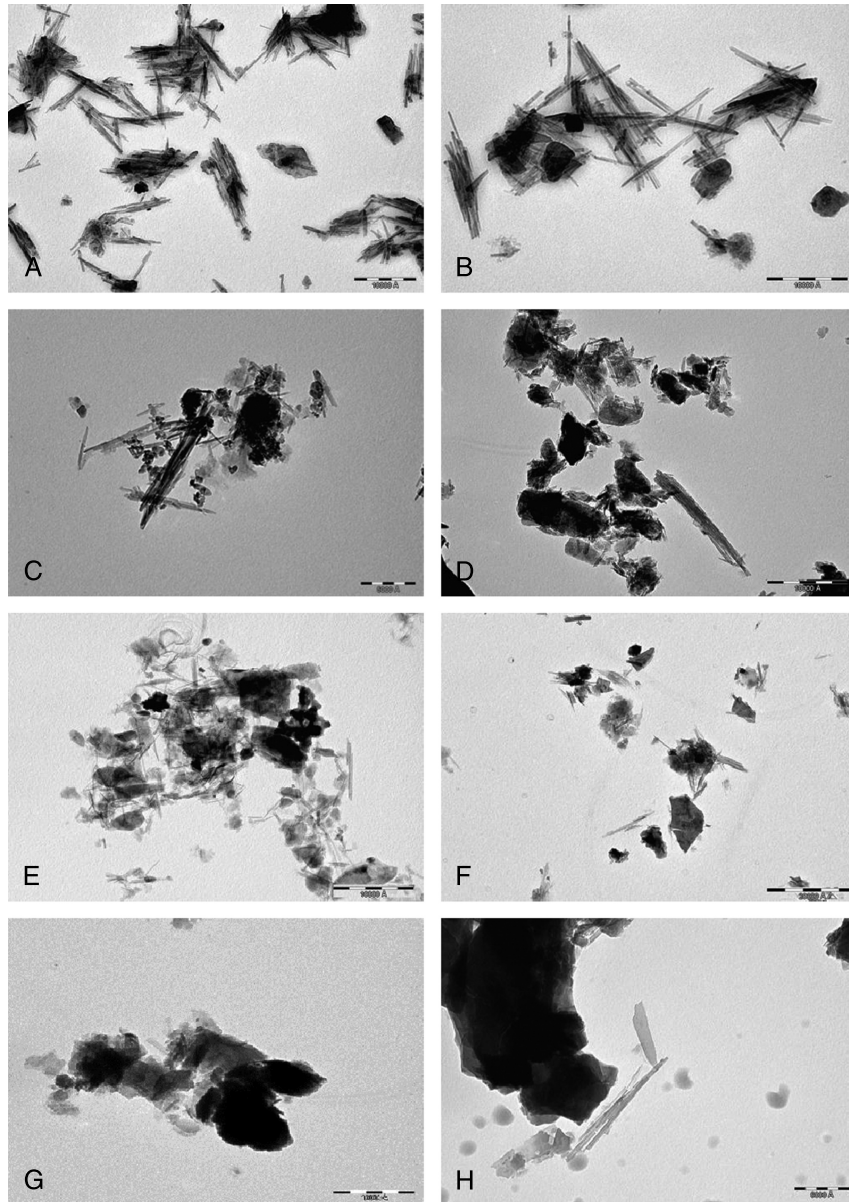
Neof ormation of palygorskite seems to need large activities of Si and Mg with a pH of about 8 (Singer, 1989). Khademi and Mermut (1998) discussed three ideal conditions for palygorskite formation in soils and sediments of central Iran: (i) increase in Mg/Ca ratio caused by gypsum crystallization in the shallow water bodies; (ii) increasing environmental pH caused by the warm climate; and (iii) increasing soluble Si caused by enriched hydrothermal solutions.

Observed typical morphology of palygorskite (Fig. 9) demonstrated the existence of it in all soils and sedimentary rocks, although the quantities varied in different samples. The long well-preserved fibers in gypsic horizons of soil 16 (Fig. 9A) and soil 10 (Fig. 9B) revealed considerable levels of palygorskite. Significant quantities of palygorskite were found in the salic (Fig. 9C), calcic (Fig. 9D), and argillic horizons (Fig. 9E) and some epipedons (Fig. 9F), although they distributed to the next level based on frequencies and crystallites.

The amounts of palygorskite in soils were more dominant than in rock samples (Fig. 9G and H); nevertheless, rock samples indicated palygorskite. Sedimentary rocks of Fars Formations indexed more amounts. The results of the current study cannot support the possible mechanisms in soil solution for pedogenic

formation of palygorskite. However, the amount of palygorskite and also its morphology are different in the studied horizons compared with the parent material of the studied soils.

In addition, calcareous, saline, and gypsiferous soils can provide buffered alkaline media with necessary anions and cations for palygorskite crystallization, but characteristics of the solution chemistry of the gypsiferous soils may provide a more favorable medium for this purpose, as noted by Owliaie et al. (2006). The occurrence of palygorskite in parent rocks as well as broken fibers in the soil surface may be the result of inheritance or of a detrital origin, as proposed in former investigations of arid soils in Iran (Khademi and Mermut, 1998; Hojati et al., 2012). Therefore, detrital (from parent rocks), transformation, and neof ormation of palygorskite are possible contemporary forms of it in soils. Khormali and Abtahi (2003) indicated that there is an inverse correlation between palygorskite and smectite with regard to the soil-available moisture, as expressed by  $P/ET^{\circ}$  (ratio of mean annual precipitation to mean annual reference crop evapotranspiration). At  $P/ET^{\circ}$  values greater than 0.4, palygorskite transforms to smectite. Therefore, it can be concluded that the critical available moisture as expressed by  $P/ET^{\circ}$  is approximately 0.4, above which palygorskite is highly unstable and transforms mainly to smectite.



**FIG. 9.** TEM micrographs of representative clay samples in Gypsic horizon (A and B), Salic horizon (C), Calcic horizon (D), Argillic horizon (E), Parent rocks (G and H), demonstrating the presence of palygorskite in all samples but with different morphology and frequency.

In the present study, calculated  $P/ET^{\circ}$  values were less than 0.4 (Table 1), so it can be concluded that palygorskite is stable and may not transform to smectite.

The ideal chemical formula of palygorskite according to the Bradley (1940) model is  $Si_8Mg_5O_{20}(OH)_2(OH)_4 \cdot 4H_2O$ . Martin-Vivaldi and Robertson (1971) questioned the usefulness of an ideal formula, although the five positions cannot be filled and a dioctahedral mineral with a structural formula  $Si_8O_{20}Al_2Mg_2(OH)_2(OH)_4 \cdot 4H_2O$  is accepted. For palygorskite, Drits and Sokolova (1971) established that the sum of octahedral cations for bulk analyses ranges from 3.45 to 4.33 with (Al + Fe) between 1.12 and 2.3 for five octahedral positions. Paquet et al. (1987) studied 145 individual particles from palygorskite-smectite and sepiolite-smectite assemblages and affirmed that the octahedral composition fields of the smectites and fibrous clays partly overlap.

The palygorskite field is both in the dioctahedral as well as between the trioctahedral and dioctahedral domains.

Suárez et al. (2007) and Garcia-Romero and Suarez (2010) concluded, based on the octahedral composition, that palygorskite could be of four different types. (i) Ideal palygorskite, with an octahedral composition near the ideal palygorskite, similar contents of Al and Mg, and negligible substitutions. (ii) Common palygorskite, where the Al content is less than in the ideal formula and, as a consequence, the Mg content is greater, but the number of octahedral cations is close to 4 (vacant octahedral positions = 1). Although Al may be partially substituted by Fe(III) and/or Mg, this type of palygorskite has dioctahedral character. (iii) Magnesian palygorskite is the most trioctahedral extreme. The number of octahedral cations is more than 4 (vacant octahedral positions = 1). Palygorskite can be Fe rich, a category of Fe-rich palygorskite

could be used when  $Fe > Al$ . (iv) Aluminic palygorskite, the most dioctahedral extreme, which contains palygorskites with a total number of octahedral cations (p.h.u.c.) of less than 4, with  $R3/R2 > 1$  and  $Mg < 2$ .

Taking into account the structural formulae proposed for the palygorskites from the present study (Table 5), samples belong to type II, type III, and type IV, although most palygorskites studied here correspond to type III and aluminic palygorskites (type IV). A category of Fe-rich palygorskite could be used when  $Fe > Al$  in the different types described above. According to this classification, Fe-rich palygorskite samples (N.1 and NE.8) could be classified as Fe-rich type III. Al palygorskite, common palygorskite, Mg-palygorskite, and Fe-palygorskite also exist in the samples. Among these, the smallest Al content corresponds to NW5, the most trioctahedral palygorskite; and the highest Al content belongs to E.12, the most dioctahedral palygorskite found in the present study. Palygorskite can have a degree of tetrahedral substitution from almost 23% (NE.8) to less than 2%. Garcia-Romero and Suarez (2010) studied 21 samples of palygorskite from Spain and indicated that palygorskite occupies the most aluminic-magnesian and dioctahedral extreme.

## Rock

Rock samples representing all the profiles were analyzed by both morphological/micromorphological techniques and by powder XRD analysis. Palygorskite, smectite, IM, and kaolinite were present in clay-size particles of parent rock samples (Table 4). Rock samples from the eastern parts consisted of only chlorite and illite in the clay fraction. Smectite varies from less than 2 to 40% (average, 14%); this mineral is best represented between north and south.

Carbonates, gypsum, quartz, chlorite, mica (biotite/muscovite), amphiboles, and apatite were detected as major rock minerals based on XRD studies. Thin sections were used along with the XRD studies for more detailed investigation of rock minerals. Petrographic studies revealed carbonate rocks, fossiliferous carbonate rocks (Fig. 10A), mudstone (Fig. 10B), gypsum (Fig. 10C), dolomite (Fig. 10D), and sandstone (Fig. 10F), as dominant sedimentary

rocks in thin sections. Carbonate rocks formed under variable paleogeographic (lagoon, intertidal, coastal, and external shelf) conditions. Sandstones contained quartz, feldspars, muscovite, biotite, apatite, and fragments of sedimentary and metasedimentary rocks with silicic (Fig. 10E), carbonatic, and gypsic cementing agents.

Thin sections of sandstones indicate that their mineralogy is dominated by quartz, with variable amounts of feldspars and rock fragments, suggesting derivation mainly from acid igneous rocks, and older sandstones with mainly silicic and carbonate cements. Angular, coarse, and poorly sorted grains of siliciclastic sedimentary rocks (sandstones) are signs of immature sandstones where energy of deposition has decreased (Blatt et al., 1980).

The concentration of mica (muscovite and biotite) in the rocks is dominated by biotite (Fig. 9F). The average mica contents in rock and soil analyzed samples are approximately 30 and 25%, respectively. The locally high concentrations of mica reached approximately 40% (Fig. 9F) such as sandstones (Aghajari Formations) along the southern region.

The total contents of easily weatherable minerals (feldspars and biotite) are usually high in rocks. Weathering and soil formation have not affected the composition of the sand and silt fraction in most of the investigated soils. Biotite is the only mineral that showed clear signs of weathering in most of the soils. The rapid weathering of biotite is believed to be caused by the seasonal alternation of strong reduction and following oxidation resulting in the liberation of iron in the ferrous form and precipitation of free iron oxides.

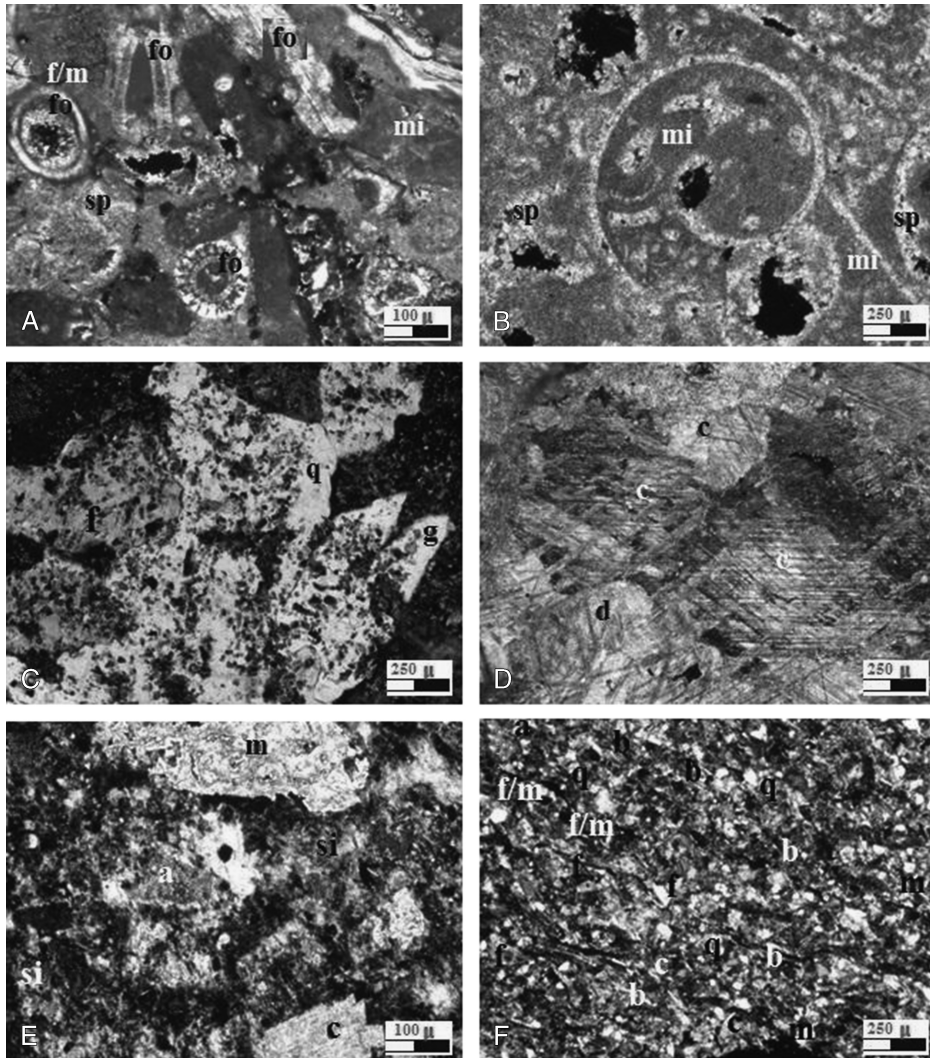
## Provenance

Carbonate precipitation has been observed in soils that are frequently subjected to evaporative concentration of solutions (Mahjoory, 1979; Kohut and Dudas, 1995). When precipitation occurs, the remaining water is enriched in salts, which precipitate when their concentration in water becomes high enough that they can no longer remain in solution. The less soluble a mineral is, the more readily it precipitates out of solution. Calcium carbonates (calcite) and dolomite are among the least soluble evaporate minerals, so these minerals precipitate first. Calcite generally contains

TABLE 5. Structural Formulae of Palygorskite

| Soil No. | Palygorskite Formulae  | $\Sigma_{oc}^*$    | Type |
|----------|--|--------------------|------|
| N.1      | (Si <sub>6.94</sub> Al <sub>1.06</sub> ) [Mg <sub>2.61</sub> Al <sub>0.87</sub> Fe <sub>0.88</sub> ] O <sub>20</sub> (OH) <sub>2</sub> (OH <sub>2</sub> ) <sub>4</sub> K <sub>0.37</sub> Ca <sub>0.1</sub> Na <sub>0</sub>     | $\Sigma_{oc}=4.36$ | III  |
| N.2      | (Si <sub>7.57</sub> Al <sub>0.43</sub> ) [Mg <sub>2.53</sub> Al <sub>1.23</sub> Fe <sub>0.50</sub> ] O <sub>20</sub> (OH) <sub>2</sub> (OH <sub>2</sub> ) <sub>4</sub> K <sub>0</sub> Ca <sub>0.03</sub> Na <sub>0.21</sub>    | $\Sigma_{oc}=4.26$ | III  |
| N.3      | (Si <sub>7.9</sub> Al <sub>0.10</sub> ) [Mg <sub>2.29</sub> Al <sub>1.20</sub> Fe <sub>0.56</sub> ] O <sub>20</sub> (OH) <sub>2</sub> (OH <sub>2</sub> ) <sub>4</sub> K <sub>0.1</sub> Ca <sub>0</sub> Na <sub>0.32</sub>      | $\Sigma_{oc}=4.05$ | III  |
| NW.4     | (Si <sub>7.9</sub> Al <sub>0.10</sub> ) [Mg <sub>4.02</sub> Al <sub>0.34</sub> Fe <sub>0.20</sub> ] O <sub>20</sub> (OH) <sub>2</sub> (OH <sub>2</sub> ) <sub>4</sub> K <sub>0.03</sub> Ca <sub>0.17</sub> Na <sub>0.3</sub>   | $\Sigma_{oc}=4.56$ | III  |
| NW.5     | (Si <sub>8</sub> ) [Mg <sub>4.1</sub> Al <sub>0.30</sub> Fe <sub>0.17</sub> ] O <sub>20</sub> (OH) <sub>2</sub> (OH <sub>2</sub> ) <sub>4</sub> K <sub>0</sub> Ca <sub>0.13</sub> Na <sub>0.24</sub>                           | $\Sigma_{oc}=4.57$ | III  |
| NW.6     | (Si <sub>7.8</sub> Al <sub>0.20</sub> ) [Mg <sub>3.32</sub> Al <sub>0.74</sub> Fe <sub>0.34</sub> ] O <sub>20</sub> (OH) <sub>2</sub> (OH <sub>2</sub> ) <sub>4</sub> K <sub>0.2</sub> Ca <sub>0</sub> Na <sub>0.17</sub>      | $\Sigma_{oc}=4.4$  | III  |
| NW.7     | (Si <sub>8</sub> ) [Mg <sub>2.42</sub> Al <sub>1.03</sub> Fe <sub>0.33</sub> ] O <sub>20</sub> (OH) <sub>2</sub> (OH <sub>2</sub> ) <sub>4</sub> K <sub>0.07</sub> Ca <sub>0.17</sub> Na <sub>0.1</sub>                        | $\Sigma_{oc}=3.78$ | II   |
| NE.8     | (Si <sub>6.13</sub> Al <sub>1.87</sub> ) [Mg <sub>2.80</sub> Al <sub>0.65</sub> Fe <sub>0.98</sub> ] O <sub>20</sub> (OH) <sub>2</sub> (OH <sub>2</sub> ) <sub>4</sub> K <sub>0.1</sub> Ca <sub>0</sub> Na <sub>0.43</sub>     | $\Sigma_{oc}=4.43$ | III  |
| NE.9     | (Si <sub>7.80</sub> Al <sub>0.20</sub> ) [Mg <sub>1.69</sub> Al <sub>1.52</sub> Fe <sub>0.63</sub> ] O <sub>20</sub> (OH) <sub>2</sub> (OH <sub>2</sub> ) <sub>4</sub> K <sub>0.20</sub> Ca <sub>0.07</sub> Na <sub>0.07</sub> | $\Sigma_{oc}=3.57$ | IV   |
| E.10     | (Si <sub>7.96</sub> Al <sub>0.04</sub> ) [Mg <sub>1.72</sub> Al <sub>1.74</sub> Fe <sub>0.33</sub> ] O <sub>20</sub> (OH) <sub>2</sub> (OH <sub>2</sub> ) <sub>4</sub> K <sub>0.17</sub> Ca <sub>0.07</sub> Na <sub>0.2</sub>  | $\Sigma_{oc}=3.79$ | IV   |
| E.11     | (Si <sub>7.75</sub> Al <sub>0.25</sub> ) [Mg <sub>2.3</sub> Al <sub>1.25</sub> Fe <sub>0.63</sub> ] O <sub>20</sub> (OH) <sub>2</sub> (OH <sub>2</sub> ) <sub>4</sub> K <sub>0.1</sub> Ca <sub>0</sub> Na <sub>0.37</sub>      | $\Sigma_{oc}=4.18$ | III  |
| E.12     | (Si <sub>7.68</sub> Al <sub>0.32</sub> ) [Mg <sub>2.83</sub> Al <sub>2.06</sub> Fe <sub>0.36</sub> ] O <sub>20</sub> (OH) <sub>2</sub> (OH <sub>2</sub> ) <sub>4</sub> K <sub>0.43</sub> Ca <sub>0.46</sub> Na <sub>0.03</sub> | $\Sigma_{oc}=3.25$ | IV   |
| W.13     | (Si <sub>7.35</sub> Al <sub>0.65</sub> ) [Mg <sub>2.35</sub> Al <sub>0.99</sub> Fe <sub>0.77</sub> ] O <sub>20</sub> (OH) <sub>2</sub> (OH <sub>2</sub> ) <sub>4</sub> K <sub>0.20</sub> Ca <sub>0.30</sub> Na <sub>0.13</sub> | $\Sigma_{oc}=4.29$ | III  |
| W.14     | (Si <sub>7.76</sub> Al <sub>0.24</sub> ) [Mg <sub>1.6</sub> Al <sub>1.69</sub> Fe <sub>0.56</sub> ] O <sub>20</sub> (OH) <sub>2</sub> (OH <sub>2</sub> ) <sub>4</sub> K <sub>0.2</sub> Ca <sub>0.43</sub> Na <sub>0.1</sub>    | $\Sigma_{oc}=3.85$ | IV   |
| S.15     | (Si <sub>7.92</sub> Al <sub>0.08</sub> ) [Mg <sub>2.00</sub> Al <sub>1.38</sub> Fe <sub>0.56</sub> ] O <sub>20</sub> (OH) <sub>2</sub> (OH <sub>2</sub> ) <sub>4</sub> K <sub>0.07</sub> Ca <sub>0.1</sub> Na <sub>0.1</sub>   | $\Sigma_{oc}=3.94$ | II   |
| PR.16    | (Si <sub>7.68</sub> Al <sub>0.32</sub> ) [Mg <sub>1.13</sub> Al <sub>1.51</sub> Fe <sub>0.40</sub> ] O <sub>20</sub> (OH) <sub>2</sub> (OH <sub>2</sub> ) <sub>4</sub> K <sub>0</sub> Ca <sub>0.17</sub> Na <sub>0.04</sub>    | $\Sigma_{oc}=3.04$ | IV   |
| PR.17    | (Si <sub>7.74</sub> Al <sub>0.26</sub> ) [Mg <sub>1.36</sub> Al <sub>1.82</sub> Fe <sub>0.56</sub> ] O <sub>20</sub> (OH) <sub>2</sub> (OH <sub>2</sub> ) <sub>4</sub> K <sub>0.03</sub> Ca <sub>0.26</sub> Na <sub>0.07</sub> | $\Sigma_{oc}=3.74$ | IV   |

\*Summation of octahedral cations.



**FIG. 10.** Thin sections of fossiliferous calcite (A), mudstone (B), gypsic (C), dolomite (D), and sandstone (E, F) rocks samples from Miocene sediments. f/m: Fe/Mn impregnation or nodules; fo: fossil; mi: micritic; sp: sparitic; f: feldspars; c: carbonate; d: dolomite; m: muscovite; a: amphiboles; si: silicic; b: biotite. Stoops (2003).

less than 5% mol of  $MgCO_3$  because it usually precipitates from more diluted solutions, which typically have low Mg/Ca ratios. However, this ratio increases as calcite formation continues, allowing the subsequent precipitation of Mg-carbonates such as magnesite and dolomite (Eugster and Hardie, 1978; Boettinger and Richardson, 2001).

Higher relative amounts of carbonate and especially dolomite were found in the Zagros structure, with a decreasing trend from north to south and west to east, addressed to early outgoing of north formations. In general, this observation suggests the earlier uplifting of the Zagros zone than the Makran zone from the water that surrounded the study area in the second and third geologic eras.

Abundance of kaolinite in marine sediments suggests a warm and humid climate with high rainfall in the Tethys region during the early Paleocene. The gradual disappearance of kaolinite, giving way to palygorskite and sepiolite, suggests the progressive development of arid climatic conditions in parts of the Tethys from the late Paleocene to the early Eocene, which has continued to the current periods. In most carbonatic rocks of the Zagros setting, neomorphism of carbonate mud (micrite) to sparry carbonate

(sparite) cement (Fig. 10B) demonstrates evolution toward a more pacific, warmer, and shallower environment.

## CONCLUSIONS

Because the region is located in a warm and dry climate, most of the soil limitations are related to saline or gypsfiferous materials, lack of drainage, unevenness, low depth, loam soil type, and coarse structure. In general, the farther from the mountainous area to the valley and lowlands and southern regions, the moderation of weather lessens and heat increases.

Parent materials greatly influence soil development and the distribution of soils on the southern Iran that are generally alkaline in nature and high in soluble salts and calcium contents. Characteristically, these soils are very low in OM/humus. Most of the element reserve is present in unweathered mineral forms.

The sand and silt fractions of these soils have mixed mineralogy, that is, silt is composed of quartz, calcite, alkali feldspars, plagioclase, dolomite, 14 Å-phyllsilicate, and mica minerals, whereas its sand fraction contained gypsum in addition to the minerals present in silt. The main minerals in silt and sand

fractions are quartz and alkali-feldspars and in total analyzed soil samples are calcite and dolomite.

Illite, chlorite, and palygorskite were present throughout the clay fraction (<2 mm) of soil and rock samples in most profiles. Smectite was more abundant in soils developed in the Zagros zone, decreasing from west to east. Vermiculite was absent, whereas kaolinite was present although in minor amounts in rare series. The clay mineralogy results showed that inheritance is likely the main source of kaolinite, smectite, chlorite, and illite, whereas *in situ* neof ormation during the Tertiary of shallow saline and alkaline environment could be the dominant cause of palygorskite presence in the sedimentary rocks.

Mineralogical variations were mostly associated with nature and composition of parent materials and degree of weathering. In the arid homogeneous climate of southern Iran, where sedimentary covering is sparse and lithological variability is pronounced, parent rock has a significant role on weathering and mineral distribution. The chemical decomposition of feldspars, biotite, and amphiboles is relatively slow, with the result that the weathering products have a higher sand/silt plus clay ratio than in more humid climates.

## REFERENCES

- Abtahi, A. 1977. Effect of saline and alkaline ground water on soil genesis in semiarid southern Iran. *Soil Sci. Soc. Am. J.* 41:583–588.
- Abtahi, A. 1985. Synthesis of sepiolite at room temperature from SiO<sub>2</sub> and MgCl<sub>2</sub> solution. *Clay Miner.* 20(4):521–523.
- Abu-Zeid, M. M., A. R. Baghdady, and H. A. El-Etr. 2001. Textural attributes, mineralogy and provenance of sand dune fields in the greater Al Ain area, United Arab Emirates. *J. Arid Environ.* 48(4):475–499.
- Aghanabati, A. 2004. Geology of Iran, 586 [in Persian]. Geological Survey of Iran, Iran.
- Allen, B. L., and B. F. Hajek. 1989. Mineral occurrence in soil environments. *In: Minerals in Soil Environments*, 2nd Ed. J. B. Dixon, S. B. Weed (eds.). Soil Science Society of America, Madison, WI, pp. 199–278.
- Alon, B. G., T. Alon, and T. Z. Noemi. 2006. The sustainability of arid agriculture: Trends and challenges. *Ann. Arid Zone* 45(2):1–31.
- Amelung, W., W. Zech, X. Zhang, R. F. Follett, H. Tessen, E. Knox, and K. W. Flach. 1998. Carbon, nitrogen, and sulfur, pools in particle-size fractions as influenced by climate. *Soil Sci. Soc. Am. J.* 62(1):172–181.
- Amundson, R. G., O. A. Chadwick, J. M. Sowers, and H. E. Doner. 1989. The stable isotope chemistry of pedogenic carbonates at Kyle Canyon, Nevada. *Soil Sci. Soc. Am. J.* 53:201–210.
- Azaroff, L. V., and M. J. Buerger. 1958. *The Powder Method in X-Ray Crystallography*. McGraw-Hill Book Co., New York, NY.
- Azizi, P., S. Mahmoodi, H. Torabi, M. H. Masihabadi, and M. Homae. 2011. Morphological, physico-chemical and clay mineralogy investigation on gypsiferous soils in southern of Tehran, Iran. *Middle-East J. Sci. Res.* 7(2):153–161.
- Balba, A. M. 1995. Management of soils problem in arid ecosystems. CRC Press, Boca Raton, FL, pp. 250.
- Banaei, M. H., M. Bybordi, A. Moameni, and M. J. Malakouti. 2005. The Soils of Iran: New Achievements in Perception, Management and Use [In Persian]. Agricultural Research and Education Organization and Soil and Water Research Institute, Tehran, Iran, pp. 482.
- Bashour, I., and A. Sayegh. 2007. *Methods of Analysis for Soils in Arid and Semi-Arid Regions*. FAO, Roma, p. 119.
- Berner, E. K., and R. A. Berner. 1996. *Global Environment—Water, Air, and Geochemical Cycles*. Prentice Hall, NJ.
- Birkeland, P. W. 1999. *Soils and Geomorphology*, 3rd Ed. Oxford University Press, New York, NY.
- Blatt, H. G. Middleton, and R. Murray. 1980. *Origin of Sedimentary Rocks*. Prentice Hill, NJ, p. 781.
- Boettinger, J. L., and J. L. Richardson. 2001. Saline and wet soils of wetlands in dry climates. *In: Wetland Soils: Genesis, Hydrology, Landscapes, and Classification*. J. L. Richardson and M. J. Vepraskas (eds.). Lewis Publishers, New York, NY, pp. 383–390.
- Bouyoucos, G. J. 1962. Hydrometer method improved for making particle size analysis of soils. *Agron. J.* 54:464–465.
- Bradley, W. F. 1940. The structural scheme of attapulgite. *Am. Miner.* 25:405–411.
- Brady, N. C. 1990. *The Nature and the Properties of Soils*. Macmillan Publishing Company, New York, NY, pp. 621.
- Bullock, P., N. Federoff, A. Jongerius, G. Stoops, T. Tursina, and U. Babel. 1985. *Handbook for Soil Thin Section Description*. Waine Research Publication, Wolverhampton, UK, pp. 152.
- Chapman, H. D. 1965. Cation exchange capacity. *In: Methods of Soil Analysis*. C. A. Black (ed.). Part 2. American Society of Agronomy, Madison, WI, pp. 891–901.
- Choopani, S., M. M. Khosroshahi, M. Gholampoor, and K. Mirakhorlo. 2006. Determination of desert areas of Hormozgan Province from geological aspect [Abstract in English]. *Iranian J. Range Desert Res.* 1 (22):17–26.
- Drits, V. A., and G. V. Sokolova. 1971. Structure of palygorskite. *Soviet Phys. Crystallogr.* 16:288–231.
- Dultz, S. 2000. Effects of silicate weathering and lessivage on K-content in forest soils derived from Pleistocene sediments. *J. Plant Nutr. Soil Sci.* 163(3):299–305.
- Eugster, H. P. and L. A. Hardie. 1978. Saline lakes. *In: Lakes: Chemistry, Geology and Physics*. A. Lerman (ed.). Springer-Verlag, Berlin, Germany, pp. 237–294.
- Fitzsimmons, K. E., J. W. Magee, and K. J. Amos. 2009. Characterization of aeolian sediments from the Strzelecki and Tirari Deserts, Australia: Implications for reconstructing palaeoenvironmental conditions. *Sediment. Geol.* 218:61–73.
- Garcia-Romero, E., and M. Suarez. 2010. On the chemical composition of sepiolite and palygorskite. *Clays Clay Miner.* 58(1):1–20.
- Gharaee, H. A., and R. A. Mahjoory. 1984. Characteristics and geomorphic relationships of some representative Aridisols in southern Iran. *Soil Sci. Soc. Am. J.* 48:115–119.
- González, C., I. Valverde, and A. L. Lafuente. 2007. Mineralogical and geochemical characterization of a diapiric formation in the North of Spain. *Catena* 70(3):375–387.
- Graham, R. C., and A. T. O'Geen. 2010. Soil mineralogy trends in California landscapes. *Geoderma* 154:418–437.
- Hojati, S., H. Khademi, A. Faz Cano, and A. Ahmad Landi. 2012. Characteristics of dust deposited along a transect between central Iran and the Zagros Mountains. *Catena* 88(1):27–36.
- Jackson, M. L. 1979. *Soil Chemical Analysis Advanced Course*. 2nd ed. M. L. Jackson, Madison, WI, p. 895.
- Johnsson, M. J., 1993. The system controlling the composition of clastic sediments. *In: Processes Controlling the Composition of Classic Sediments*. M. J. Johnsson and A. Basu (eds.). Boulder, Colo, Geological Society of America, pp. 1–19.
- Khademi, H., and A. Mermut. 1998. Source of Palygorskite in gypsiferous Aridisols and associated sediments from Central Iran. *Clay Miner.* 33 (4):561–575.
- Khademi, H., and J. M. Arocena. 2008. Kaolinite formation from palygorskite and sepiolite rhizosphere soils. *Clays Clay Miner.* 56(4):429–436.
- Khormali, F., and A. Abtahi. 2003. Origin and distribution of clay minerals in calcareous arid and semi-arid soils of Fars Province, southern Iran. *Clay Miner.* 38:511–527.



- Khormali, F. A. Abtahi, S. Mahmoodi, and G. Stoops. 2003. Argillic horizon development in calcareous soils of arid and semi-arid regions of southern Iran. *Catena* 53:273–301.
- Kohut, C. K., and M. J. Dudas. 1995. Evaporite mineralogy and trace-element content of salt-affected soil in Alberta. *Can. J. Soil Sci.* 73:399–409.
- Konyuhov, A. I., and B. Maleki. 2006. The Persian Gulf Basin: Geological history, sedimentary formations, and petroleum potential. *Lithol. Miner. Resour.* 41(4):344–361.
- Mahjoory, R. A. 1979. The nature and genesis of some salt-affected soils in Iran. *Soil Sci. Soc. Am. J.* 43:1019–1024.
- Martin-Vivaldi, J. L., and R. H. S. Robertson. 1971. Palygorskite and sepiolite (the hormites). *In: The Electron Optical Investigation of Clays.* J. A. Gard (ed.). Mineralogical Society, London, UK. pp. 225–275.
- Mashhady, A. S., M. Reda, H. J. Wilson, and R. C. Mackenzie. 1980. Clay and silt mineralogy of some soils from Qasim. *Saudi Arab. J. Soil Sci.* 31(1):101–115.
- Moore, D. M., and R. C. Reynolds Jr. 1997. X-Ray Diffraction and the Identification and Analysis of Clay Minerals. 2nd Ed. Oxford University Press, New York, NY.
- Murashkina, M. A., R. J. Southard, and G. S. Pettygrove. 2007. Silt and fine sand fractions dominate K fixation in soils derived from granitic alluvium of the San Joaquin Valley, California. *Geoderma* 141:283–293.
- Myriam, M., M. Suarez, and J. M. Martin-Pozas. 1998. Structural modifications of palygorskite and sepiolite under acid treatment. *Clays Clay Miner.* 46(3):225–231.
- Nelson, D. W., and L. E. Sommers. 1996. Total carbon, organic carbon, and organic matter. *In: Methods of Soil Analysis, Part 2.* 2nd Ed. A. L. Page et al. (eds.). American Society of Agronomy, Madison, WI. pp. 961–1010.
- Nettleton, W. D., and F. F. Peterson. 1983. Aridisols. *In: Pedogenesis and Soil Taxonomy. II. The Soil Orders.* L. P. Wilding, N. E. Smeck, G. F. Hall (eds.). Elsevier, Amsterdam, The Netherlands, pp. 165–215.
- NRCS. 1998. Soil Quality Indicators: pH. Soil Quality Information Sheet. National Soil Survey Center in cooperation with the Soil Quality Institute, USDA, Washington, DC.
- O'Geen, A. T., W. A. Hobson, R. A. Dahlgren, and D. B. Kelly. 2008. Evaluations of soil properties and hydric soil indicators for vernal pool catenas in California. *Soil Sci. Soc. Am. J.* 72:727–740.
- Owliaie, H. R., A. Abtahi, and R. J. Heck. 2006. Pedogenesis and clay mineralogical investigation of soils formed on gypsiferous and calcareous materials, on transect, southwestern Iran. *Geoderma* 134:62–81.
- Page, A. L., W. D. Burge, T. J. Ganje, and M. J. Garber. 1967. Potassium and ammonium fixation by vermiculitic soils. *Soil Sci. Soc. Am. J.* 31:337–341.
- Paquet, H., J. Duplay, M. M. Valleron-Blanc and G. Millot. 1987. Octahedral compositions of individual particles in smectite-palygorskite and smectite-sepiolite assemblages. Proceedings of the International Clay Conference, Denver, 1985. *In: L. G. Schultz, H. Van Olphen, and F. A. Mumpton (eds.). The Clay Minerals Society, Bloomington, IN.* pp. 73–77.
- Parry, S. A., M. E. Hodson, E. H. Oelkers, and S. J. Kemp. 2011. Is silt the most influential soil grain size fraction? *Appl. Geochem.* 26:S119–S122.
- Pye, K. 1986. Mineralogical and textural controls on the weathering of granitoid rocks. *Catena* 13:47–57.
- Rasmussen, C., M. S. Torn, and R. J. Southard. 2005. Mineral assemblage and aggregates control carbon dynamics in a California forest. *Soil Sci. Soc. Am. J.* 69:1711–1721.
- Reheis, M. C. 2006. A 16-year record of eolian dust in southern Nevada and California, USA: Controls on dust generation and accumulation. *J. Arid Environ.* 67:487–520.
- Reid, D. A., R. C. Graham, R. J. Southard, and C. Amrhein. 1993. Slickspot soil genesis in the Carrizo Plain, California. *Soil Sci. Soc. Am. J.* 57:162–168.
- Ruffell, A. H., G. D. Price, J. Mutterlose, K. Kessels, E. Baraboshkin, and D. R. Grocke. 2002. Paleochimate indicators (clay minerals, calcareous nannofossils, stable isotopes) compared from two successions in the late Jurassic of the Volga Basin (SE Russia). *Geol. J.* 37:17–33.
- Salinity Laboratory Staff. 1954. Diagnosis and Improvement of Saline and Alkali Soils. Agriculture Handbook. vol. 60. US Department Agriculture, Washington, DC.
- Sharland, P. R., R. Archer, D. M. Casey, R. B. Davies, S. H. Hall, A. P. Heward, A. D. Horbury, and M. D. Simmons. 2001. Arabian plate sequence stratigraphy. *GeoArabia* 2:371.
- Singer, A., W. Kirsten, and C. Bühmann. 1995. Fibrous clay minerals in the soils of Namaqualand, South Africa: Characteristics and formation. *Geoderma* 66(1):43–70.
- Singer, A. 1989. Palygorskite and sepiolite group minerals. *In: Minerals in Soil Environments.* J. B. Dixon, S. B. Weed (eds.). Soil Science Society of America, Madison, WI. pp. 829–872.
- Snyder, C. 2007. Efficient Fertilizer Use—Soil pH Management. Peachtree Corners, Ga, International Plant Nutrition Institute, p 20.
- Soil Conservation Service. 1972. Soil Survey Laboratory Methods and Procedures for Collecting Soil Samples. Rev. Ed. USDA-SCS, Soil Survey Investigation Report No. 1. US Government Printing Office, Washington, DC.
- Soil Survey Staff. 1999. Soil Taxonomy, US Department of Agriculture and Natural Resources Conservation Service, Agriculture Handbook 436, 2nd ed. US Department of Agriculture, Natural Resources Conservation Service, Washington, DC.
- Soil Survey Staff. 2010. Keys to Soil Taxonomy, 11th Ed. US Department of Agriculture, Natural Resources Conservation Service, Washington, DC.
- Southard, A. R., and R. W. Miller. 1966. Parent material—Clay relations in some northern Utah soils. *Proc. Soil Sci. Soc. Am. J.* 30:97–101.
- Stam, M. 2004. Soil as significant evidence in a sexual assault/attempted homicide case. *In: Forensic Geoscience: Principles, Techniques, and Applications.* Special Publication, vol. 232. K. Pye, D. J. Croft (eds.). Geological Society, London, UK, pp. 295–299.
- Stoops, G. 2003. Guidelines for the Analysis and Description of Soil and Regolith Thin Sections. Soil Science Society of America, Madison, WI, p. 182.
- Suárez, M., E. García-Romero, M. Sánchez del Río, P. Martinetto, and E. Dooryhée. 2007. The effect of octahedral cations on the dimensions of the palygorskite cell. *Clay Miner.* 42:287–297.
- Sultan, K. 2006. Clay mineralogy of Central Victorian (Creswick) soils: Clay mineral contents as a possible tool of environmental indicator. *Soil Sediment Contam.* 15(4):339–356.
- Taylor, G., and R. A. Eggleton. 2001. Regolith Geology and Geomorphology. John Wiley and Sons, Chichester, UK.
- Twidale, C. R. 1990. The origin and implications of some erosional landforms. *J. Geol.* 98:343–364.
- Ugolini, F. C., S. Hillier, G. Certini, and M. J. Wilson. 2008. The contribution of aeolian material to an Aridisol from southern Jordan as revealed by mineralogical analysis. *J. Arid Environ.* 72(8):1431–1447.
- UNESCO. 1977. Map of World Distribution of Arid Regions. Centre National de la Recherche Scientifique, Paris, France.
- Van der Merwe, G. M. E., M. C. Laker, and C. Buhmann. 2002. Clay mineral association in melanic soils of South Africa. *Austr. J. Soil Res.* 40(1):115–126.
- Van Wambeke, A. R. 2000. The Newhall Simulation Model for Estimating Soil Moisture and Temperature Regimes. Department of Crop and Soil Sciences, Cornell University, Ithaca, NY.
- Wilson, M. J. 1999. The origin and formation of clay minerals in soils: Past, present and the future perspectives. *Clay Miner.* 34(1):7–25.
- Woodruff, L. G., W. F. Cannon, D. D. Eberl, D. B. Smith, J. E. Kilburn, J. D. Horton, R. G. Garrett, and R. A. Klassen. 2009. Continental-scale patterns in soil geochemistry and mineralogy: Results from two transects across the United States and Canada. *Appl. Geochem.* 24(8):1369–1381.
- Ziegler, M. 2001. Late Permian to Holocene paleofacies evolution of the Arabian Plate and its hydrocarbon occurrences. *GeoArabia* 6(3):445–504.