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Pleistocene glaciation in Kurdistan

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With 17 figures and 3 tables.

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Zusammenfassung. Die Berge Kurdistans sind ein Teil des Taurus-Zagros-Gebirgszuges, der sich durch die südliche Türkei, den nördlichen Irak und den südwestlichen Iran hinzieht und das anatolisch-iranische Hochland vom mesopotamischen Tiefland trennt. Der Kamm erreicht Höhen von 3000-4000 m im Abschnitt des Cilo Dagh-Gebietes in der südöstlichen Türkei bis zum 250 km entfernten Gebiet des Algurd Dagh im nördlichen Iran. In südöstlicher Richtung senkt sich die Kammhöhe auf 2200 bis 2800 m, erreicht jedoch im Zardeh Kuh des südlichen Iran örtlich wieder Höhen von 4000 m. Die höchsten Erhebungen des Gebirges liegen gewöhnlich im Gürtel der metamorphen Gesteine. Die äußeren Kämme bauen sich meist aus langen Faltenzügen mesozoischer Kalkgesteine auf; in den Vorbergen sind es Faltenzüge, die sich aus Sedimenten bis hinauf zum Pliozän zusammensetzen. Innerhalb Kurdistans werden die einzelnen Gebirgszüge von 4 Hauptzuflüssen des Tigris (Khabur, großer und kleiner Zab und Diyala) durchschnitten.

Das Klima Kurdistans ist durch winterliche Niederschläge und sommerliche Dürre gekennzeichnet. Die Regenmengen werden teilweise von Zyklonen gebracht, die vom Mittelmeer herüberziehen, teilweise aber auch durch Umströmung einer Antizylone, deren Zentrum im Winter über der arabischen Halbinsel liegt. Die regionalen Niederschläge nehmen mit der Höhe zu und betragen von 300 mm pro Jahr in den äußeren Vorbergen bis über 1000 mm in den höchsten Teilen des Gebirges im Gebiet des Cilo Dagh und Algurd Dagh. Die Niederschlagsmenge erreicht hier nicht nur ein Maximum, weil die Berge hoch und massig sind, sondern auch deshalb, weil das Streichen der Ketten von E nach SE umschwenkt. Die Stürme, welche der äußeren Flanke in östlicher Richtung vom Mittelmeer her folgen, werden gezwungen, über das Gebirge zu steigen oder sie werden nach SE abgelenkt. Jenseits des Gebirges, auf den Hochflächen Anatoliens und des Irans, nimmt die Regenmenge auf 300 bis 500 mm ab.

Die Vegetation Kurdistans bezeichnet recht deutlich die verschiedenen Klimazonen. Die untere Baumgrenze liegt gewöhnlich zwischen 700 und 1000 m auf den äußeren Vorbergen der Ketten und begleitet in dieser Höhe fast das gesamte Gebirge. Sie folgt ungefähr der 500-mm-Niederschlagslinie. Das Waldland besteht vorwiegend aus Eichen; dieses ist örtlich infolge von Holzeinschlag und Ziegenfraß allerdings nur noch Strauchwerk. In den höheren Teilen der Waldzone tritt gelegentlich Wacholder auf. Ahorn, Walnuß, Weißdorn, Mandel und Esche kommen zusammen mit Eiche in mittleren Höhenlagen vor. Pistazien und Olivenbäume finden sich an einigen trockeneren Stellen. Die obere Baumgrenze auf den äußeren Ketten zieht sich in einer Höhe von ungefähr 2000 m hin. Sie ist in dieser Höhe wahrscheinlich bedingt durch die Temperatur (Januarmittel ca. 10° C). In SE-Richtung, dem Gebirge entlang im Iran, wo die höchsten Erhebungen gewöhnlich unter 3000 m liegen, reicht die Waldbedeckung weiter ins Landesinnere und hört an

einer „inneren Baumgrenze“ auf, wo die jährliche Niederschlagsmenge ca. 500 mm und die Höhenlage etwa 1300-1500 m beträgt.

Glaziale Erscheinungen des Pleistozäns wurden hauptsächlich in 3 Regionen untersucht: im Gebiet des Algurd Dagh im Irak, dem nahe gelegenen Ruwandiz-Flußgebiet und dem Cilo-Dagh-Gebiet in der Türkei. Die Kämme in der Nähe des Algurd Dagh erreichen Höhen von 3000-3500 m. Sie liegen zum Entwässerungssystem so, daß sich ausgedehnte pleistozäne Gletscher an den Nordhängen bildeten, die ihre Zungen durch enge Schluchten südwärts in die Nebentäler des großen Zab bis auf Höhen von 1100 m hinab vorschoben. Die oberen Teile von breiten Tälern wurden durch Glazialschutt verstopft. Ausgeprägte Moränengürtel haben sich nicht gebildet; Seen und andere kleine Eintiefungen sind vorhanden. Kleine Kare, deren Böden bis auf 1500 m heruntergehen, wurden auf den nach Norden zu abfallenden Hängen der Nebenketten festgestellt. Im Tal des Ruwandiz, eines der Hauptzuflüsse des großen Zab, liegen 40—60 m über dem heutigen Flußbett 30 m mächtige Terrassen, die aus fluvio-glazialen Kiesen des Pleistozäns aufgeschottert worden sind. Obwohl auf einigen vom Ruwandiz durchquerten Kämmen frische Kare entdeckt wurden, enden die Terrassen nicht in einem ausgeprägten Moränen-Komplex, und es ist daher möglich, daß sie älter sind als der letzte Hauptvorstoß des Eises. Die Beziehungen werden kompliziert durch Ablagerungen, die Bergstürzen im Quellgebiet zugeschrieben werden. In die Terrassen selbst ist Schutt und Bodenmaterial eingeschaltet; sie werden von mächtigem Schutt überlagert, der in die letzte eiszeitliche Phase wie auch in das Postglazial gehören kann. Auf frühere pleistozäne Ereignisse in diesem Gebiet weisen noch höher gelegene Bänke einer Kalksteinbreccie an den Berghängen (mit Höhlen des Moustiér), sowie der Überrest einer Ablagerung aus Sand und Kies und dünne Lagen eines limnischen Silts und Kalksteines in einer Höhe von 250 m über dem Ruwandiz. Im Norden des Cilo Dagh in der südöstlichen Türkei wurden glaziale Ablagerungen des Pleistozäns dem großen Zab entlang bis herunter auf 1500 m gefunden. Sie wurden durch Gletscher herangebracht, die ihren Ursprung auf der Nordseite des Cilo Dagh sowie auf nördlich und nordwestlich gelegenen Nebenketten hatten. Die tiefsten festgestellten Kare liegen in einer Höhe von ungefähr 1800 m, doch lagen die aufgesuchten Gebiete im nördlichen Teil des Vereisungsgebietes, wo die Niederschlagsmenge geringer ist als im eigentlichen Gebiet des Cilo Dagh und seiner äußeren Flanke.

Obwohl heute keine Gletscher im Algurd Dagh-Gebiet vorhanden sind, wurden mehrere kleine Reste von Bobek in Karen des Cilo Dagh verzeichnet. Die gegenwärtige Schneegrenze auf den Schattenseiten wird auf eine Höhe von ca. 3300 m gelegt. Pleistozäne Kare in 2100 m Höhe im Gebiet des Cilo Dagh und in 1500 m Höhe im Gebiet des Algurd Dagh fordern eine Erniedrigung der Schneegrenze im Pleistozän um 1200 bis 1800 m, eine Zahl, die wesentlich höher liegt als die von Bobek geforderten 700 m.

Wenn die pleistozäne Erniedrigung der Schneegrenze als alleinige Folge der Temperaturerniedrigung angesehen wird, muß die mittlere Jahrestemperatur um mindestens 12° C tiefer gelegen haben (bezogen auf einen vertikalen Temperaturabfall von 0,7° C auf 100 m). Ein solcher Wert wäre genau so groß, wie der für Mitteleuropa angenommene, wo Permafrost, Tundren-Flora und -Fauna offensichtlich weit verbreitet waren. In Kurdistan gibt es keine Frosterscheinungen oder paläontologischen Belege, die derartig niedrige Temperaturen andeuten würden. So ist es wahrscheinlich, daß die Vereisung die Folge sowohl stärkerer Schneefälle als auch einer geringen Temperaturabsenkung war. Dieser Schluß gilt nur für die äußere Flanke der Berge Kurdistans, die wesentlich größere winterliche Schneefälle in dem Maße erhalten haben können, wie die mediterranen Stürme am Rande einer verstärkten asiatischen Antizyklone an Intensität gewannen. Auf den Hochflächen Anatoliens und des Irans muß die Niederschlagsmenge nicht unbedingt größer gewesen sein; die Suche nach verläßlichen geologischen und paläontologischen Beweisen muß in diesem Gebiet noch fortgesetzt werden.

Trotz des Nachweises unterschiedlicher klimatischer Bedingungen in Kurdistan während der letzten Vereisungsphase des Pleistozäns gibt es wenig Beweise dafür, daß der klimatische Umschwung gegen Ende des Pleistozäns für die Entwicklung des Menschen vom Jäger zum Ackerbauer und Viehzüchter entscheidend war. Die Übergangsstadien liegen in dem Zeitraum von 11000 bis 9000 Jahren vor heute. Es ist wahrscheinlich, daß der Klimawechsel, der den Rückzug der Gletscher bewirkte, zu jener Zeit im wesentlichen abgeschlossen war. Auf jeden Fall kann der pleistozäne Klimawechsel nur eine höhenmäßige Verlagerung der Lebensbereiche innerhalb der Berge Kurdistans, der Vorberge und der mesopotamischen Rumpffläche zur Folge gehabt haben, so daß sogar während der Vereisungsperioden Gebiete vorhanden waren, die solchen Tieren und Pflanzen Lebensmöglichkeiten boten, die domestiziert werden konnten, sobald der Mensch das dafür notwendige Kulturniveau erreicht hatte.

Abstract. The mountains of Kurdistan are a portion of the Taurus-Zagros, mountain arc that extends through southern Turkey, northern Iraq, and southwestern Iran and separates the Anatolian-Iranian Plateaus from the Mesopotamian Lowland. The crest reaches elevations of 3000-4000 m in the segment from the Cilo Dagh area in southeastern Turkey for 250 km to the Algurd Dagh area in northern Iraq. Southeastward the crestal elevation descends to 2200-2800 m,

but in the Zardeh Kuh of southern Iran it again locally reaches 4000 m. The highest part of the range is generally in the belt of metamorphic rocks. The outer ridges are formed mostly by long folds of Mesozoic limestone, giving way in the foothills to folds in sediments as young as Pliocene. Within Kurdistan the ranges are cut transversely by four major tributaries of the Tigris River, namely the Khabur, Greater Zab, Lesser Zab, and Diyala Rivers.

The climate of Kurdistan is marked by winter precipitation and summer drought. The rains are brought in part by cyclonic disturbances from the Mediterranean Sea, and in part by circulation around an anticyclone centered in winter over the Arabian peninsula. The regional precipitation increases with elevation in the mountains, and ranges from about 300 mm per year in the outer foothills to more than 1000 mm in the highest part of the range in the area of the Cilo Dagh and Algurd Dagh. The precipitation reaches a maximum here not only because the mountains are high and massive but also because here the trend of the ranges shifts from east to southeast, and the storms which follow the outer flank of the range from the Mediterranean Sea eastward are forced to rise over the mountains or be diverted sharply to the southeast. Inland from the mountains on the high Anatolian and Iranian Plateaus the precipitation falls abruptly to 300-500 mm.

The vegetation in Kurdistan closely reflects the climate. The lower treeline has a general elevation of 700-1000 m on the outer foothills of the mountains for most of the distance along the ranges, and follows approximately the 500 mm precipitation line. The woodland consists dominantly of deciduous oak, locally reduced to scrub by woodcutters and goats. In addition juniper may be found generally in the upper part of the forest zone. Maple, walnut, hawthorne, almond, and ash occur with the oak at middle elevations, and pistachio and olive are found on some of the drier sites. The upper treeline on the outer ranges occurs at an elevation of about 2000 m, and is presumably limited at this elevation by the temperature (January mean about 10° C). Southeastward along the range in Iran, where the crests of the ridges are generally less than 3000 m, the forest cover extends farther inland and terminates in what may be considered an inner treeline, where the precipitation is about 500 mm and the elevation about 1300-1500 m.

Pleistocene glacial features were studied principally in three regions, the Algurd Dagh area in Iraq, Ruwandiz River area nearby, and the Cilo Dagh area in Turkey. The ridges near Algurd Dagh reach 3000-3500 m in elevation, and are so located with respect to the drainage that extensive Pleistocene glaciers formed on the northern slopes and flowed through gaps southward down valleys tributary to the Greater Zab River to elevations as low as 1100 m. The upper portions of broad valleys were plugged with glacial debris without distinct morainic loops but with lakes and other small depressions. Small cirques with floors as low as 1500 m were found on north-facing slopes of subsidiary ridges.

The Ruwandiz River, one of the major tributaries of the Greater Zab River, contains Pleistocene glacio-fluvial gravels at least 30 m thick in terraces 40-60 m above the present river. Although fresh cirques were found on some of the ridges crossed by the Ruwandiz River, the terraces do not head in a distinct moraine complex, and it is possible that they pre-date the last major glacial advance. The relations are complicated by the presence of deposits attributed to landsliding in the headwater region. There terraces themselves contain intercalated colluvium and soil; they are overlain by thick colluvium that may represent the last glacial phase as well as the post-glacial. Earlier Pleistocene events in this region are recorded by still higher benches of limestone breccia (with Mousterian caves) on the hill slopes as well as by a remnant of a deposit of sand and gravel and even thin layers of lacustrine silt and limestone standing 250 m above the Ruwandiz River.

In the area north of Cilo Dagh in southeastern Turkey, Pleistocene glacial deposits were found along the Greater Zab River as low as 1500 m elevation. They were supplied by glaciers originating on the north side of Cilo Dagh as well as on subsidiary ridges to the north and west. The lowest Pleistocene cirques identified have an elevation of about 1800 m, but the only areas visited were actually in the northern part of the glaciated area, where the precipitation is less than in the main Cilo Dagh and its outer flank.

Although no modern glaciers exist in the Algurd Dagh area, several small remnants were mapped by BOBEK in cirques in the Cilo Dagh. The modern snowline on shaded exposures is placed at about 3300 m. Pleistocene cirques at 2100 m in the Cilo Dagh area and at 1500 m in the Algurd Dagh area and at 1500 m in the Algurd Dagh area imply a Pleistocene snowline depression of 1200-1800 m, a figure much larger than the 700 m postulated by BOBEK.

If Pleistocene depression of the snowline is assumed to be a result of depression of temperature alone, then the mean annual temperature must have been at least 12° C lower (based on a vertical temperature gradient of about 0.7° C/100 m). Such a value is as great as that infered for central Europe, where permafrost and tundra flora and fauna were apparently widespread. For Kurdistan there are no frost features or paleontologic records to indicate such low temperatures, and it is probable that the glaciation was result of increased snowfall as well as moderately lower temperature. This conclusion applies only to the outer flank of the Kurdish mountains, which may have received greatly increased winter snowfall as Mediterranean storms were intensified on the margin of a strengthened Asiatic anticyclone. On the Anatolian and Iranian Plateaus, however, the precipitation may not necessarily have been greater, and search for reliable geologic and paleontologic evidence must be made in this region.

Despite the evidence for markedly different climatic conditions in Kurdistan during the last glacial phase of the Pleistocene, there is little evidence yet that climatic change near the Pleistocene was critical in the evolution of early man from hunters to farmers and herders. The transitional stages occurred during the period about 11,000 to 9,000 years ago, and it is probable that the climatic change that brought about glacier recession had been essentially by that time. In any case, the Pleistocene climatic changes may have involved only an altitudinal shifting of the life zones within the Kurdish mountains and foothills and the Mesopotamian piedmont, and that even during the glacial period there were natural habitats suitable for the animals and plants destined to be domesticated as soon as man reached the requisite cultural level.

Introduction

Kurdistan is a mountain land where the countries of Iraq, Turkey, and Iran join together. High, rugged ridges of the great Taurus-Zagros mountain arc, 3,000-4,000 m above sea level, extend eastward from the Mediterranean and thence southeastward to the Persian Gulf, separating the Mesopotamian Lowland from the Anatolian-Iranian Plateaus (Fig. 1). These mountains divide southwestern Asia geologically, physiographically, climatically, and culturally - - they have long served as an effective barrier to cultural interchange between Mesopotamia and the interior plateaus, but at the same time have provided summer pasturage for nomadic groups from the lowlands and secure strongholds for the Kurdish mountain people.

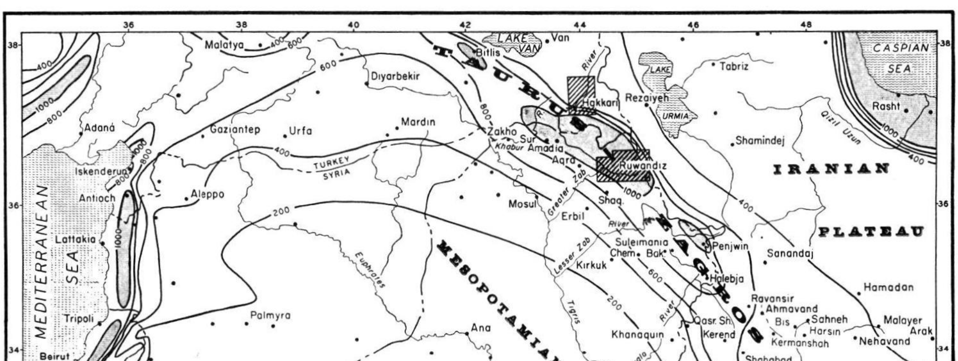

Fig. 1. Map of eastern Mediterranean region showing how precipitation is controlled by elevation and by proximity to the Mediteranean Sea. Data for Iran from GANJI (1960), for northern Iraq from DENNIS (1953), for Turkey from LEMBKE (1940), and for the rest of area from FISH & DUBERTRET (1946).

The growing evidence that the foothills and piedmont of the Kurdish mountains served as one principal locus for the beginnings of village life and the domestication of plants and animals (BRAIDWOOD, HOWE, *et al.,* 1960) draws attention to the past physical and climatic environments that may have influenced this important cultural transformation. Although it seems reasonable that the Pleistocene and post-glacial climatic changes known to have occurred in Europe should be reflected in some manner in the Near East, the proof of such changes must come from studies in the field. A very sensitive and at the same time a well-recorded climatic indicator is glaciation, so a first step in reconstruction of paleoclimate should be the study of Pleistocene glaciers in nearby mountains. BOBEK (1940) had reported expanded Pleistocene glaciers in the Cilo Dagh area of Turkish Kurdistan, DEMORGAN (1909, p. 92) had referred to glacial features in the Zagros ridges of Iranian Kurdistan and adjacent Luristan, but the geologists of the Iraq Petroleum Company, who have traveled extensively in the high mountains of Iraqi Kurdistan, never reported any glacial features in this region. The present paper records new evidence for the extent of glaciation in these mountains, and reviews the older evidence.

Opportunities to study briefly the mountains of Kurdistan came to the writer in 1951, 1954-55, and 1960 while he was attached to the prehistoric projects of the Oriental Institute (University of Chicago). He is indebted to the John Simon Guggenheim Memorial Foundation and the Wenner-Gren Foundation for Anthropological Research for fellowships, to R. J. BRAIDWOOD for providing supplemental funds from the Oriental Institute and the National Science Foundation and for organizing base facilites and other indispensable liaison, to colleagues on the projects for cooperation and assistance, to H. V. Dunnington and R. V. Browne of the Iraq Petroleum Company in Kirkuk for courtesies and geological advice, to Leo Anderson and Charles Simkins of the Khuzestan Development Service, Ahwaz, for loan of facilities and equipment, and to countless government officialis and villagers for hospitalities and other aids.

Physical and climatic setting

The crest of the Taurus-Zagros Mountains reaches elevations of more than 4,000 m above sea level in the Cilo Dagh in southeasternmost Turkey near the headwaters of the Grater Zab River. The crest remains high at 3,000-3,500 m. eastward along the Iraq-Iran frontier in Kurdistan for several km, then lowers to 2200-2800 m in the headwater region of the Lesser Zab and Diyala Rivers. Still farther to the southeast in the Zardeh Kuh of Iran the crest again locally attains elevations of 4,000 m, and then dies out in southern Iran.

The axis of the range consists largely of Paleozoic and Mesozoic metamorphic and volcanic rocks thrust outward (to the southwest) over linear folds of Mesozoic limestone and Tertiary terrestrial deposits. The folds range from symmetrical gently-plunging structures to steep-flanked flat-topped features etched by stream erosion and weathering to rugged ridges and precipitous gorges. The folds decrease in intensity outward in the foothill zone, and as the Mesopotamian Lowland is approached they are confined to gentle structures in the Tertiary rocks.

The topographic forms closely reflect the geologic stratigraphy and structure. The metamorphic and volcanic zone consists of rugged nonlinear ridges reaching elevations of 3800-4100 m. The belt of folds is marked by several topographic forms: long sweeping ridges and valleys made of successive hard and soft beds on the flanks of folds, broadly sloping anticlinal ridges giving way along the strike to sharp paired hogbacks where the fold crest is breached, and minor synclinal ridges. Important stratigraphic units expressed topographically in alternate ridges and valleys are (1) resistant massive Triassic and Jurassic limestones, (2) soft Lower Cretaceous marls (3) resistant Upper Cretaceous limestones, (4) generally soft Upper Cretaceous and Eocene thin-bedded limestones, marls, red shales, and conglomerates, (5) resistant Eocene limestones (6) soft Miocene redbeds, and (7) Pliocene siltstones (soft) and conglomerates (locally resistant). Complications in these generalized relations are introduced by lateral facies changes in bedding of limestone, for example, and these are reflected in the topography by reduction in prominence of a particular ridge or valley.

The stratigraphic sequence and the structural relations indicate that uplift of the mountain belt started as early as the Eocene, and the onceextensive seas of this area and the Mesopotamian Lowland became restricted to allow formation of evaporite deposits. Accelerated uplift during the Tertiary is recorded by terrestrial Miocene and Pliocene deposits, locally thousands of feet thick; culmination of the orogeny produced not only the Pliocene conglomerates but also the expansion of the deformed zone so that the late Tertiary sediments themselves were folded. Just how long the deformation persisted is not known, but it is possible that some of the river terraces and other Pleistocene landforms owe their genesis to continued crustal movements.

The master drainage of Kurdistan consists of four large transverse streams which head in the high mountains or the interior plateaus and flow transversely across the fold ridges to join the Tigris River along the axis of the Mesopotamian Lowland. These four streams -- the Khabur, Greater Zab, Lesser Zab, and Diyala -- have probably inherited their transverse courses from their Tertiary equivalents that deposited the late Tertiary piedmont sediments, and are thus essentially antecedent streams. As the individual anticlinal folds rose as potential barriers, the streams maintained their transverse courses by cutting sharp canyons, and repeated uplifted of the mountain belt periodically rejuvenated the streams. Some adjustments from such transverse courses are recorded by the short longitudinal subsequent segments, and by the fact that some of the ridge crossings are localized by plunges of the folds or by other favorable structures. There may also be some cases of autosuperposition from the less resistant younger sediments of the folded series onto the more resistant Cretaceous limestones.

The climate of the Taurus-Zagros mountain arc is typically Mediterranean but there are differences along its great length. Practically all the precipitation comes in the fall, winter, and spring. The storms are of two types (Boesch 1941). The first type is produced by cyclonic disturbances that have traversed the length of the Mediterranean or have regenerated in secondary low-pressure centers over Cyprus at the front of outbreaks of polar continental air from western Asis (El Fandy 1946; Butzer 1958, p. 22). These storms move as cold fronts and are characterized by strong winds and by snow at appropriate altitudes. They are dominant in the coastal mountains and in the western Taurus, and they invade the interior along two principal tracks, one across the "Syrian Saddle" south of the Taurus arc, the second across the north-Palestinian upland to the Hauran area south of Damascus. The invasions are common especially in the fall and spring, and the northern track occasionally carries the storms far eastward even to the Persian Gulf and Pakistan. The passage of these cold fronts of Mediterranean origin is followed by periods of clear cold weather with strong winds as the cold air from the northern plateaus breaks out over the border mountains onto the Mesopotanian piedmont.

The second type of storm is related to the Arabian anticyclone, which reaches maximum development in mid-winter and becomes connected with the Asiatic high (Boesch 1941). It serves as blocking high to the Mediterranean cyclones, but has its own precipitation pattern in the Taurus-Zagros Mountains, with gentle warm-front rains in contrast to the extreme storminess of the Mediterranean type. In the spring it breaks down and allows the passage of an increasing number of Mediterranean cyclones.

Thus with these two types of storm the mountain belt is favored by rain or snow from November to April. The gentle winter rains in particular are vital in building up the soil moisture for the spring grain crop, and the spring storms add to the snow pack in the

Table 1 Temperature and precipitation in southwestern Iran

	LAT. °	ELEV. m.	No. Years	TEMPERATURE(1) ° C. Mean Annual	Mean Jan.	Mean Ju'y	Range	PRECIPITATION(2) No. Years	Annual mm.
PIEDMONT — Steppe and Hot Desert Steppe									
Abadan	30	6	9	25	12.4	36.2	24	8	157
Ahwaz	31	20	6	25.4	12	37	25	3	242
Shushtar	32	80	8	26.4	13.5	38.7	25	8	351
Gotvand	32	110	6	25.4	12.1	37.4	25	4	379
Dizful	32	150	7	25.9	13 4	38.3	27	6	402
Mesjid-i-Suleiman	32	240	7	23.8	10	37.0	25	21	466
Qasr Sharin	34	330	3	21.9	8.6	33.6		4	382
ZAGROS OUTER RIDGES — Oak Forest									
Sefid Dasht	33	1150	4	18.6	5.3	32.1	27		
Khorramabad	34	1170	9	17.0	4.9	29.3	25	7	519
Elam	34	1400						13	733
Kerend	34	1600	6	14.6	3.1	26.8	24	4	703
ZAGROS INNER RIDGES AND IRANIAN PLATEAU — Steppe and Cold Desert Steppe									
Dorud	33	1450	3	13.6	—0.6	27.9	28		
Nehavand	34	1650	4	12.1	—0.9	24.5	25		
Malayer	34	1750						7	295
Arak	34	1750	4	13.5	—1.7	27.2	25	7	313
Hamadan	35	1880	12	11.7	—1.1	24.0	25	11	359
Dareh Takt	33	1800	4	10.0	—5.5	22.3	28		
Shahabad	34	1400						3	427
Kermanshah	34	1320	16	4.1	2.0	27.1	25	15	421
Bisitun	34	1360	3	13.4	0.7	26.5	26		
Sahneh	34	1470					23	6	397
Ravansur	34	1500	5	13.5	1.6	24.6	24	4	734
Ahmavand	34	1600	3	13.0	0.8	25.1			
Harsin	34	1700					27	4	530
Sanandaj	35	1650	12	15.3	1.3	28.6		12	537
Shamindej	36	1400						5	257
Rezaiyeh	38	1330						8	417
Tabriz	38	1360						9	283

(1) Data from Khuzestan Development Service, Ahwaz, Iran
(2) Data from GANJI (1960).

Table 2 Precipitation in northern Iraq

	LATITUDE °N.	ELEVATION m.	PRECIPITATION(1) Years No.	Annual mm.	DJF %	NDJFM %
ZAGROS PIEDMONT AND MESOPOTAMIA — Steppe and Hot Desert Steppe						
Baghdad	33	35	16	154	50	80
Diwaniya	32	18	23	233	55	78
Khanaquin	34	210	14	327	55	87
Mosul	36	245	29	382	55	81
Kirkuk	35	380	14	371	54	85
Erbil	36	400	15	518	54	79
Chemchemal	36	700	12	631	52	77
ZAGROS OUTER RIDGES — Oak Forest						
Zakho	37	420	17	837	51	78
Amadia	37	1235	13	900	40	69
Sursank	37	915	7	1049	54	77
Aqra	37	715	15	949	54	82
Shaqlawa	36	1050	7	1035	60	84
Ruwandiz	37	1005	15	1014	55	81
Suleimania	36	850	15	734	50	75
Bakrajo	36	730	12	825	46	72
Halebja	35	720	18	827	55	82
Penjwin	36	1400	11	1339	57	84

(1) Data from DENNIS (1953)
DJF = December, January, February NDJFM = November through March.

mountains. In fact Boesch (1941) interprets three precipitation maxima -- the fall and spring related to the Mediterranean storms, and the winter to the Arabian anticyclone. The scarcity of long-run stations in this region makes such close differentiation uncertain, however. Bobek (1952, p. 70) places the mountain belt in his zone of spring/winter maximum and the piedmont in winter/spring.

Precipitation in the Taurus-Zagros Mountains and piedmont is controlled principally by the general altitude of the land area. Within the mountain country, however, the precipitation does not reflect closely the elevation of individual stations--equal amounts fall on mountain ridges and intervening valleys, as illustrated by the scatter of points on the precipitation/elevation curve for transects across the Zagros ridges in Iraq and Iran (Fig. 2). Total precipitation rises abruptly from 250-400 mm (10-16 inches) in the piedmont to perhaps as much as 1500 mm (60 in.) in the middle of the high mountains, but then decreases abruptly to 300-500 mm on the Iranian Plateau (Table 1-2; Fig. 2).

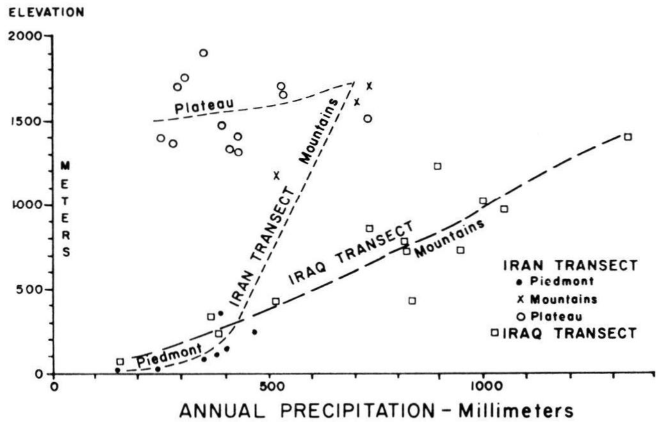

Fig. 2. Relation of precipitation to elevation in two transects across the Zagros Mountains. Data in tables 1 and 2.

The maximum precipitation in the entire Taurus-Zagros arc at present is in the high mountains of Kurdistan in the southeastern corner of Turkey and adjacent Iraq (Fig. 1). Two factors are involved here. First, the mountains in this region are higher and more massive than elsewhere along the arc, for they exceed elevations of 3,000 m for a distance of almost 250 km. They thus serve as a more effective barrier to moisture-bearing air masses moving against the mountains from Mesopotamia, whether the precipitation is associated with the passage of the Mediterranean cyclonic storms or with the occurrence of orographic or anticyclonic circulution produced by the Arabian anticyclone. The second factor, which would seem to be effective primarily with respect to the Mediterranean storms, is the rather abrupt curvature of the mountain arc just in Kurdistan. The storms which cross the Syrian saddle from the Mediterranean have a trajectory toward the northeast as they follow the flanks of the western and middle Taurus. As they reach the eastern Taurus they must be deflected to the southeast or cross the massive barrier and spread onto the Iranian Plateau. The fact that relatively few of them move very far onto the plateau is indicated by the abrupt decrease in precipitation northeast of the mountains. In his study of precipitation relations in Turkey, Louis (1944, p. 477) pointed out that the contrast between the wet side and the dry side of mountain ranges is not great when the storms travel parallel to the ranges, but when storms abut directly or obliquely against a range the contrast is sharp. Thus in Kurdistan the mountain volume and alignment are both favorable for high precipitation, with mean annual values exceeding 1200 mm in some intermontane valleys and probably 1500 mm at higher elevations.

The mountains above 1,500 m generally have continuous snow cover in winter, and the storage of this snow is an important feature in the flood regime of the large mountain rivers. Small glaciers and perennial snow patches persist in sheltered locations in the highest mountains, notably in the Cilo Dagh area in southeasternmost Turkey and in the Zardeh Kuh in Iran north of the head of the Persian Gulf. Winter temperatures vary precisely with elevation (Table 1; Fig. 3). Few villages are located within the mountains higher than 1,500 m, presumably because of the winter snow.

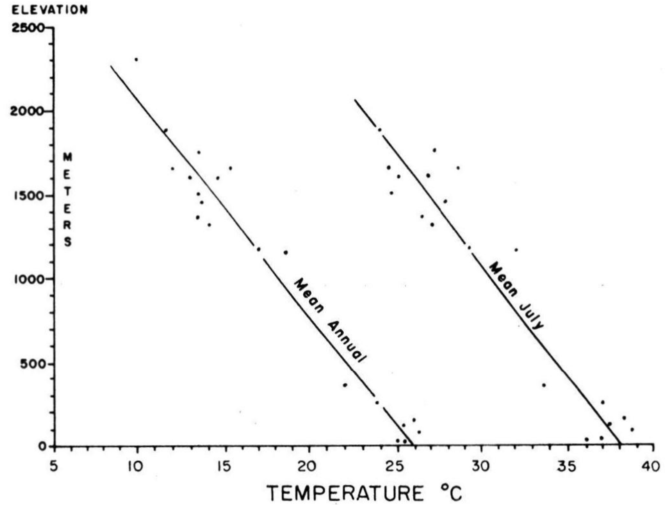

Fig. 3. Vertical air-temperature gradient for 20 ground stations in transect from southern Mesopotamia across Zagros Mountains to Iranian Plateau.

In the summer the zone of westerly cyclones moves northward and the weather in the eastern Mediterranean region is controlled largely by a low-pressure system centering in the hot lands around the Persian Gulf. The prevailing winds in Kurdistan are north and dry. Rain rarely falls in the summer, and even clouds are uncommon. Temperatures are high in the piedmont, with daily maxima generally exceeding 35° C from May to October. Above about 1,000 m, however, summer heat is less intense and nights are cool. The great range in mean temperature between July and January (about 25° C) is a reflection of the basic continentality of the climate (Table 1).

The climatic zones are closely reflected by the vegetational relations. The lower treeline has a moisture control. It rises rapidly from sea level on the Mediterranean coast to about 700 m elevation 50 km inland from the coast, as the Mediterranean humidity decreases. Eastward through the Mesopotamian piedmont for at least 900 km the lower treeline remains at 700-1000 m elevation, and marks an annual precipitation of about 500 mm. Local variations in the elevation of the treeline are caused by special factors. It is higher on hot south-facing slopes than on north-facing slopes, where the sun is not so bright. Small valleys at low elevation within the higher mountains or close to the mountains may be forested because of the greater precipitation engendered by the nearby mountains.

Oak (mostly *Quercus persica)* dominates the forest throughout the length of the mountains. Close to the Mediterranean below 1000 m elevation, where the winters are mild, a separate zone of cold-sensitive types like *Pinus brutia,* olive *(Olea),* and various hardwood shrubs occur below the main oak forest. Farther east maple *(Acer),* hawthorne *(Crataegus),* ash *(Fraxinus),* and almond *(Amygdalus)* are minor components in the oak forest, especially in the lower part. Pistachio *(Pistacia)* is co-dominant on certain dry sites, where fig *(Ficus)* also occurs. Black pine *(P. nigra)* and juniper *(J. excelsa)* are common in the main forest zone along with oak in the western part of the Taurus Range, especially at higher elevations (LOUIS 1939, p. 96). Eastwards towards Kurdistan pine drops out; juniper extends barely into Iraq on the outer flanks of the Zagros ridges, but on the inner margin of the inner margin of the forest is found farther east on the Iranian Plateau (BOBEK 1951, p. 32).

Where the forest is undisturbed and well developed, as in the mountains above Elam, the arboreal cover exceeds 50% over broad areas, but close to villages, trails, and migration routes the oak especially has been cut for charcoal, house beams, and summer shelters, and any seeding and sprouting have been inhibited by the grazing of goats. Oak propagates by root shoots and therefore does not require full growth to the flowering stage to assure regeneration. It therefore has probably not been so easy for man to modify its gross distribution as in the case of other trees. Nonetheless, the lower treeline along the front of the Kurdish mountains next to the intensively inhabited piedmont steppe may have been raised considerably by these various disturbances, which have been practiced for millenia. The occurrence of isolated trees far below the present treeline in exposed

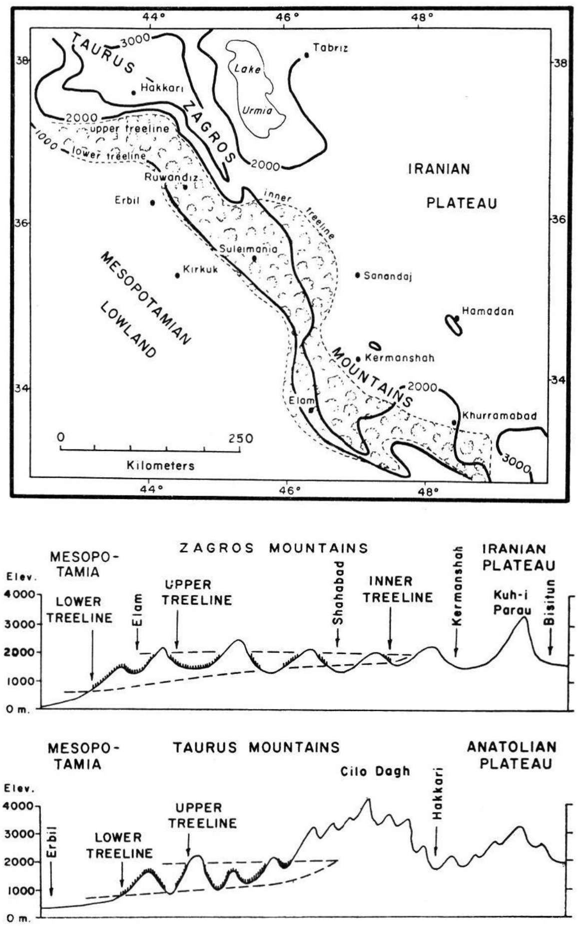

Fig. 4. Generalized map and cross-sections to show relation of treelines to elevation and geographic position in the Taurus-Zagros Mountains and adjacent regions.

localities protected by some ceremonial tradition implies more extensive woodlands in the past.

The upper treeline on the outer ridges has an elevation of about 2,000 m along the entire length of the Taurus-Zagros mountain arc from the Mediterranean to southern Iran. Its position is probably controlled by winter temperature, for there is no indication that moisture is deficient at this elevation. The precipitation at this treeline is probably at least 1,200 mm. Above the treeline is an alpine zone of herbs and low shrubs.

In addition to the lower and upper treelines on the outer Zagros ridges, there is an inner treeline that reflects primarily the decrease in moisture on the lee (northeast) side of the border mountains and on the Iranian Plateau. This treeline, like the lower treeline on the outer ridges, follows roughly the 500 mm precipitation line (Fig. 4). Its position depends primarily on distance from the axis of the high montains. Where the Zagros ridges are not high enough to have a continuous alpine zone, the inner treeline occurs about 75 km inland from the outer flank of the range, as in the segment from Kermanshah to Khorramabad in Iran. The elevation of the inner treeline in this sector is about 1,300-1,500 m. Where the main mountain mass is entirely above treeline, as in the Cilo Dagh-Algurd Dagh area under special consideration in this paper, the cold-dry plateau steppe essentially merges with the cold-moist alpine zone, and the two cannot be easily distinguished. These cold steppes are characterized by the great frequency of perennial herbs, whereas the hot steppe of the Mesopotamian piedmont has such hot summers that many perennial herbs cannot survive, and the relatively short spring growing season favors annual herbs (Pabot 1960).

Algurd Dagh area

The crest of the Zagros Range along the Iraq-Iran frontier northwest of the Ruwandiz River includes several ridges and summits reaching elevations of 3,000-3,500 m above sea level. Among these are Algurd Dagh, Hawarju Kuh, and Siah Kuh (Fig. 5, 6). Three valleys leading to the west from the crest drain eventually into the Ruwanditz River, which is a principal tributary of the Greater Zab River. These three valleys are occupied respectively by the three principal villages Bola, Birkim, and Beni, and they will be described under these names.

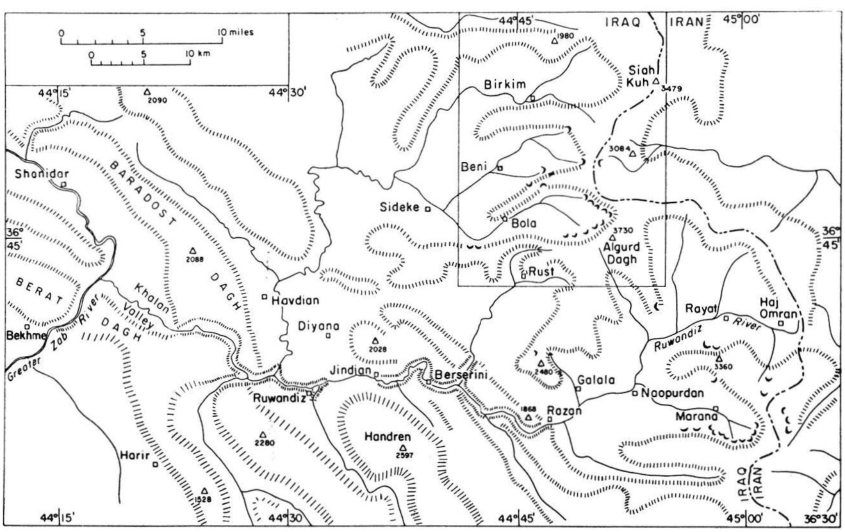

Fig. 5. Map of Algurd Dagh area and Ruwandiz River area in Zagros Mountains, Iraq-Iran. Small crescents show Pleistocene glcial cirques. Area of Figure 6 outlined by rectangle.

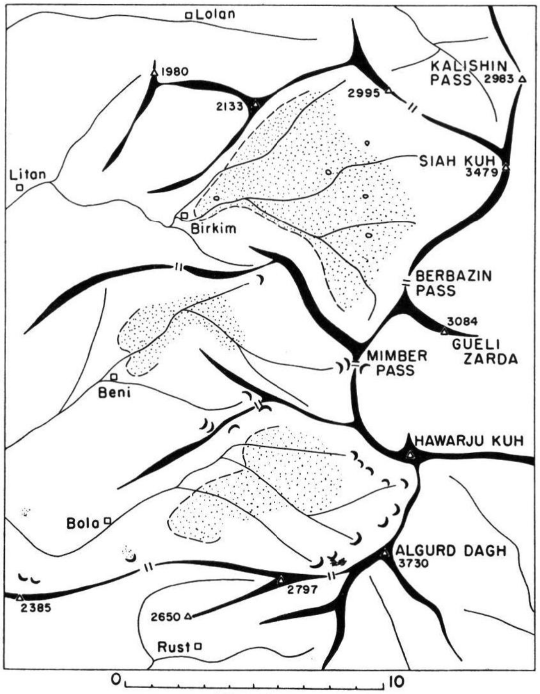

Fig. 6. Map of Algurd Dagh area, Iraq-Iran, showing distribution of glacial deposits (stipple) and cirques (small crescents).

Bola Valley

The village of Bola is located at an elevation of about 1500 m above sea level just outside the mouth of a broad basin-like valley head that is filled with glacial deposits. The stream that drains the basin enters a sharp gorge about 100 m deep just upstream from Bola, and irrigation canals are led from the head of the gorge to water the rice fields on the terrace on which the village is located. The basin is about 5 km broad and is known by the local people as Horner. The elevation of its floor rises from about 1500 m near its mouth to about 2500 m at the inner edge; then begins the steeper slope leading to the crest. The upper tree limit in this region is about 1800 m.

The basin floor has an irregular rolling topography with few rock outcrops--typical of glacial deposits. One small lake about 100 m in diameter is located at an elevation of about 2,500 m near the base of the south wall of the basin, and another lake is reported closer to the head wall. Many meadows in flat depressions suggest the former existence of other undrained depressions. The material composing the basin floor consists of glacial till, and includes fragments mostly of metamorphic rock derived from the basin rim--schist, slate, grit, limestone, volcanic breccia, red siltstone, etc.

Numerous small cirques are located on the headwall of the Horner basin below Algurd Dagh and Hawarju Kuh at estimated elevations of 2,700 m, and glaciers from these cirques presumably fed the main ice mass that must have filled the Horner basin. Other cirques can be found at lower elevations in the Bola Valley, however, away from the headwall region. For example, on the south-facing slope of Bardi Spi, which is the ridge on the north side of Bola Valley, four small cirques were identified at elevations of 1700, 2000, and 2100 m. On the opposite side of Bola valley two cirques are located at

1700 m and a third, near the pass over the ridge to the Rust Valley, has a floor at about 1450 m. These cirques are considered to be the work of small cliff glaciers, and their positions give an indication of the elevation of the Pleistocene snowline.

No distinct looped moraines could be recognized in Horner basin, except at the very head of the valley where a tongue of young drift with apparently fresher features originated in the sheltered recess north of Algurd Dagh and extended for about 4 km down into the center of the basin to an elevation of about 2000 m.

Birkim Valley

The village of Birkim is located at a comparable position to Bola in a valley farther north. The village is perched on a terrace approximately at the entrance of the broadly expanded glaciated head of the valley. Its elevation is about 1400 m above sealevel, close to the upper limit of year-round habitation in this area. Downstream form the village the valley is narrow and is cut in bedrock. The expanded valley uptream is called Birkima, and serves as excellent pasture area for nomadic and semi-nomadic groups of sheep-, goat-, and horse-breeders. Small fields are cultivated above the village to about 2100 m.

Birkima is floored by irregular glacial drift. No distinct moraines are evident, but there appear to be three breaks in the level of the floor that might represent successive ice-front positions above about 2400 m elevation. The topography is particularly hummocky. Two lakes occur at about 2400 m. One of them, called Dindara Lake, has a diameter of about 200 m. It suddenly overflowed a few years ago: its outlet stream cut a deep channel, exposing 6 m of drift, and then built a large boulder fan below. A rather steep 150 m drop leads from about 2400 m down to a less hummocky zone, which descends gradually to about 1900 m. Lakes in this area were discovered at elevations of 2400 and 2100 m in addition to many meadows. Patches of outwash gravel were found as high as 2100 m. Another steep drop leads to the final zone, which descends more gradually to about 1500 m and contains small ponds or undrained depressions at 1700 and 1550 m. Below 1500 m the surface again drops rapidly to the mouth of Birkima near the village, where the river exposes a 10 m section of till at an elevation of about 1300 m.

The color of the till is variable and reflects the local bedrock--gray slate and sandstone, green volcanic rocks, red chert and slate, gray limestone and diorite. Large knobs project above the till surface, but the regional structure is so complex that it is difficult to determine whether these knobs are bedrock or simply large till blocks. The gray till at about 2400 m elevation is oxidized to a depth of about 1 m.

Beni Valley

The village of Beni, as is the case with the other two valleys, is located approximately at the point where the broad valley head becomes constricted. It has an elevation of about 1500 m, and probobly represents the approximate limit of continuous glacial deposits. Although there are no lakes or deep exposures of till in this broad valley head, a thick filling of glacial deposits is suggested by the lack of bedrock ledges on the floor, by the rolling topography, and by the cirques on the ridges bordering the valley head.

Sideke Area

Beni valley and Bola valley descend approximately 5 km from the respective villages and join near the larger village of Sideke, which is located on a 30 m terrace at an elevation of about 950 m above sea level. Other terraces in the area are as great as 60 m above the present river. They contain coarse gravels, including boulders 1 m across. In consideration of the facts that the Sideke area is a great distance from the high mountains and that large streams of competence sufficient to carry such large boulders no longer exist here, it seems likely that the gravels are glacio-fluvial, and that the ice front stood close by. This conclusion is supported by the occurrence of till at about 1100 m near the village of Birte about 3 km upstream (southeast) from Sideke on a low ridge just above the terraces.

Ruwandiz River area

In addition to the Pleistocene glaciers of the Algurd Dagh area, several other centers were located elsewhere in the high mountains of northeastern Iraq but were not carefully studied or mapped. Most of these are in the drainage area of the Ruwandiz River, one of the principal left-bank tributaries of the Greater Zab. The Ruwandiz River heads near the Iranian frontier among the high ridges that locally reach elevations of 3000-3500 m above sea level. These ridges contain small fresh cirques in sheltered locations, but the associated glacial deposits are not well displayed in the valleys below, and it seems likely that the most recent glaciers did not fill these valleys. Downstream the river leaves the metamorphic zone and crosses successively lower ridges of the fold zone; none of these ridges shows clear signs of glaciation.

Galala Valley

A distinct cirque occurs at an elevation of about 1900 m in a 2400-3000 m ridge at the head of Galala valley, which joins the right bank of the Ruwandiz about 4 km above the fold zone. A possible cirque exists on the north-facing slope of this valley at about 1400 m. Near the mouth of the valley is a coarse gravel terrace which is probably glaciofluvial.

Marana Valley

On the opposite (left) side of the Ruwandiz River is a tributary draining west past the villages of Marana and Naopurdan. This tributary valley is bordered on the south by a high ridge (Chia-i-Mandau or Sakri Sakran) at 3000-3500 m. At least 13 cirques are

Fig. 7. Moraines at the head of Marana Valley.

located high on the sheltered northern slope of this ridge; some contain linear moraines or rock glaciers which stream down rugged slopes to an estimated elevation as low as 2000 m (Fig. 7). The northern flank of Marana Valley leads up to a 2100-2400 m ridge; here the topography is confused because of the irregular structure of limestone, but one deep cirque-like form with several small lakes occurs with floor at an estimated elevation of 1800 m. The north flank and floor of Marana Valley itself is covered with till-like debris cemented in the upper few feet to resistant breccia. Below about 1800 m the floor becomes deeply entrenched, and near the village of Marana more than 30 m of the encrusted debris are exposed along the stream. Downstream the former valley floor persists as shoulders about 150 m above the present Marana River, and has an elevation of about 1200 m near the junction of Marana valley with the Uuwandiz valley.

The debris on the flanks and floor of Marana valley is interpreted as pure glacial till, although it is recognized that alternatively the material may have originated as landslide debris from the limestone ridges to the north. The bedrock of Marana Valley itself is not well enough exposed to discern lithology favorable for landsliding. The heavy cementation of the debris and the smoothness of its topographic forms suggests that if this debris is till it may represent a stage of glaciation before the last, although the abundance of calcareous ground water and soil water on the north flank of the valley, coupled with the summer aridity, might provide conditions conducive to crust formation in a relatively short period.

Ruwandiz Headwaters

The Ruwandiz River Valley itself contains terraces and other features of probable or possible glacial origin. This river heads in the frontier region above Rayat, flows across the metamorphic zone to Razan, thence across several limestone fold ridges through the Berserini Gorge to the edge of the Diyana Plain, which is localized by the broad outcrop of Gretaceous shale. The river then enters the Ruwandiz Gorge as it obliquely crosses the next limestone anticline, and then follows the next syncline westward for many miles before joining the Greater Zab River at the entrance to Bekhme Gorge.

Rayat has an elevation of about 1100 m above sea level, and the Ruwandiz River Valley above Rayat is open in much the same way as the valleys above Birkim, Bola, and Beni described previously. The open slopes, however, are more smoothly graded and are not marked by the irregular glacial knobs and meadows which characterize those other valleys. Phyllite crops out at many localities, but the cover of debris contains many erratic stones and a thin soil. Bedrock lithology and structure are not conducive to landsliding extensive enough to explain the broad valley opening, and it seems likely that the main erosion was accomplished by glaciation. The relatively smooth grading of the slopes, however, suggests that the glaciation was older than that recorded in the Birkim and Bola valley heads. The frost climate of the last glaciation may have led to smoothing of the slopes and stripping of the soil. The ridges close to the Rayat area reach elevations of only 2100-3000 m and may not have supported glaciers large enough during the last glacial phase to fill the Rayat area. Small cirques on these ridges in sheltered localions, however, may be referred to the last glaciation. One on the ridge 3 km west of Rayat, with eastern expousure, has floor at about 1500 m. The sharp 2400 m ridge south of Rayat (i. e. north of Marana Valley) has several small cirques with northern exposure at 2100 m and a main cirque at about 1500 m. The moraine debris from this larger cirque actually reaches the valley floor near Rayat.

Other cirques are visible on the frontier ridges east of Rayat, particularly clustered on northern exposures on the 2700-3000 m ridges near the head of Marana Valley, but the drainage from these cirques leads to the Iranian side rather than to the Ruwandiz River. The low pass at the frontier at the head of the Rayat area has an elevation of only about 1775 m. The frontier ridge 8-10 km north of Rayat (3000-3300 m) was not examined, but it is doubtful that large glaciers formed on its southern slopes during the last glacial phase of the Pleistocene.

The village of Rayat itself is located opposite a nicely exposed terrace at the junction of the two main tributaries that form the Rayat opening. The terrace is about 30 m high and contains coarse boulders and poorly sorted sediment that appears to be glacio-fluvial. A few km above Rayat the terrace is undissected, and its surface grades upward laterally to the footslopes of the ridges. If the terrace gravel is the deposit of an older glaciation, it may thus be overlapped on the flanks by slope debris carried by solifluction during the last glacial phase.

Marana-Galala-Razan Segment

Downstream from Rayat the Ruwandiz River traverses a narrow segment for about 10 km in which no terraces are preserved.

Below this narrow segment the valley opens as the broad entrance of the Marana Valley is approached on the left bank. The Ruwandiz River is crowded to the right by two thick masses of debris emanating from the end of the ridge north of Marana Valley. Each fan-like protrusion is about 3 km broad as it meets the Ruwandiz. Together they have occasioned a sharp descent of about 100 m in the river gradient here along its toe (Fig. 8, 9). The surfaces of the protrusions have an elevation of about 1100 m near the river, and appear from the distance to have an irregular topography. Exposures in several river cuts at least 6 m deep show a greenish-gray clayey till-like material. At places this rests on dissected terrace gravels of the main river, and at one exposure 2 km upstream from the entrance of the Galala River the clayey "till" is separated from the underlying

Fig. 8. Ruwandiz River upstream from entrance of Marana River (foreground). Note two lobes of landslide debris that have pushed river to the right bank.

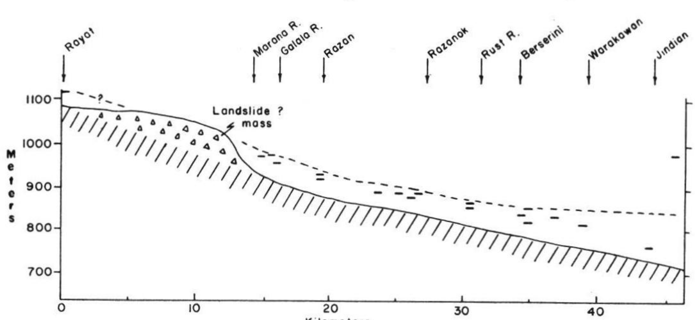

Fig. 9. Profile of Ruwandiz River from Rayat to Diyana Plain, showing height of terrace remnants and relation to Marana landslide lobe. Dotted line shows inferred profile for main glaciofluvial terrace. Elevations taken from British Indian Survey maps and by altimeter.

river gravel and its associated colluvium by a red soil. The clayey "till" is overlain in turn by younger soil and colluvium.

These two protrusions are tentatively interpreted as landslide masses rather than glacial debris, on the basis of three lines of evidence.

(1) The scars in the limestone ridge to which they may be traced face unsheltered to the south and have an elevation of only about 1800 m. True cirques of comparable elevation in this area all occur in sheltered exposures facing north or northeast.

(2) A portion of the area adjacent to the limestone ridge is underlain by shale, a rock type favorable for landsliding.

(3) The terraces of the Ruwandiz River in this area seem not to be outwash terraces graded form these till-like masses; rather they are older, and were dissected before being overridden by the protrusion. The debris masses thus were relatively dry (like landslides but unlike glaciers) and did not supply enough water and sediment for valley alluviation.

Interpretation of the relations in this area near the mouth of Marana Valley is complicated by the occurrence of another possible till within the terrace gravels. This material is light brown rather than greenish gray and is sandy rather than clayey. It was apparently derived from upstream rather than from the sedimentary rocks of Marana Valley, and is provisionally considered to be the deposit of a glacier which fed the extensive gravel terraces from this region downstream all the way to the Diyana Plain.

On the valley flanks above the river terraces in this region occur some sloping benches of resistant limestone slope-breccia. These benches rise far up the slopes as smooth surfaces, but at places are deeply dissected. In the area near Razan, just above the entrance to the Berserini Gorge, some breccia benches may be traced up to great scars on the limestone cliffs, and apparently originated as rock falls and landslides over the shale slopes that border the limestone ridges. Fragments of breccia are secondarily deposited in the terrace gravel. In the Razan area a younger landslide down into the valley bottom produced a mass of debris that contains fragments from the slope breccia, rounded cobbles from the terrace gravels, and chunks of old soil, all in a clayey matrix. Occurrences such as this, where the slippery shale bedrock is known to be present and where the ridge north of Razan is probably not high enough (with southern exposure) to have nourished glaciers, permit a fairly certain assignment to landsliding.

The geomorphic history of the Marana-Galala-Razan segment of the Ruwandiz River valley may be summarized as follows (Fig. 10):

1. Formation of slope breccia in areas of limestone, rock falls from limestone cliffs, landslides where shale occurs.

2. Entrenchment of Ruwandiz River, leaving breccia as benches.

3. Glaciation in the higher ridges of the metamorphic zone. Glaciers probably filled Marana Valley, Galala Valley, and the main Ruwandiz Valley as far downstream as the Galala region. Deposition of outwash gravels down the main stream, with some contribution from side slopes.

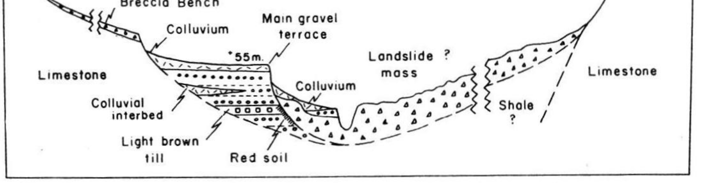

Fig. 10. Generalized composite cross-section of Ruwandiz River valley near mouth of Marana Velley showing inferred relation of Marana landslide to main glacio-fluvial terrace.

4. Entrenchment of Ruwandiz River downstream from Marana area, leaving the outwash fill as a terrace about 40 m high.

5. Second glaciation, confined largely to cirques in sheltered locations but as low as 1400 m. Landslides reached mouth of Marana Valley from limestone ridge to east and blocked Ruwandiz River.

6. Late stage of glaciation caused small cirques at 2700-3000 m high on the walls of larger cirques on Chia-i-Mandau south of Marana Valley.

Berserini Gorge

The Ruwandiz River in the segment of the Berserini Gorge through part of the limestone fold belt is marked by a continuation of the gravel terraces found between Galala and Razan. The elevations of about 9 terrace remnants over a distance of about 20 km were measured; the maximum height of most patches is about 40 m above the river, although some gravels were found as high as 60 m above the river. In several cases at least 30 m of gravel is exposed, so the terraces must be considered as depositional features formed after an earlier deep cutting of the valley. The gravels consist of rounded pebbles and cobbles (and locally boulders) of metamorphic and igneous rocks from the higher ridges upstream as well as limestone from the local ridges. Interfingered with the rounded river gravels are lenses of colluvium contributed from the adjacent slopes. The terrace surfaces themselves are overlain by colluvium, which forms a smoothly graded slope up to bounding ridges.

The relative antiquity of the gravel terrace is suggested by the extensive unconformable overburden of colluvium. At one locality, 7 km upstream from the Jindian bridge, the eroded terrace gravel is overlain on the slope by colluvial rubble, a red soil, and several meters of stratified sand and silt (Fig. 11). It appears here that the terrace gravel, which reaches a maximum height of 55 m above the river, was cemented and then cliffed by partial river entrenchment. Colluvial rubble formed beneath the cliff, with contributions from black limestone bedrock as well as from the terrace conglomerate, and then red soil formed on the stabilized colluvium. Temporary damming of the gorge (perhaps by a landslide) caused deposition of sand and silt in slack water. Renewed intrenchment, with slight shift in channel position, left the silt filling perched on the valley side.

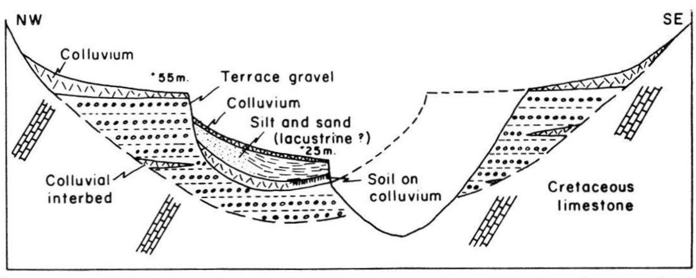

Fig. 11. Sketch cross-section across Ruwandiz River near Berserini, showing patch of lacustrine (?) deposits set in an old channel cut into glacio-fluvial terrace gravels.

Diyana Plain

Near the exit from the Berserini Gorge the terrace gravels are found at generally greater maximum heights, and above the Jindian bridge on the edge of the Diyana Plain gravel was found as much as 110 m above the river. An even more impressive remnant of stream activity is a 25 m section of sand and basal gravel resting unconformably on Cretaceous shale on the top of an isolated hill above the Ruwandiz River at the southeast end of the Diyana Plain opposite the town of Ruwandiz (Fig. 12). The top of the remnant

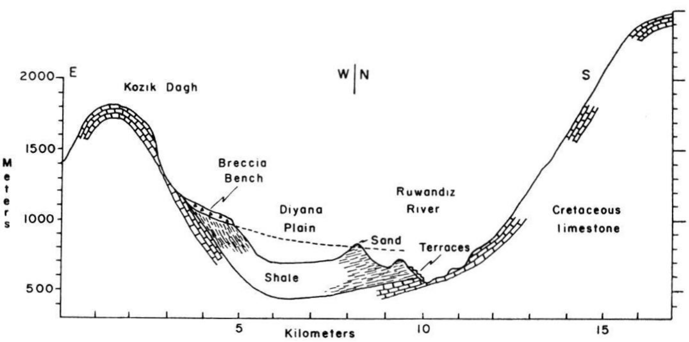

Fig. 12. Bent cross-section of Ruwandiz River gorge and Diyana Plain, showing relation of breccia bench and sand-capped hill to gravel terraces of Ruwandiz River. Elevations from British Indian Survey maps and by altimeter.

is 250 m above the river. The basal gravel contains cobbles and boulders derived from the metamorphic zone of the frontier ridges, and was certainly deposited by the Ruwandiz River. The sand contains carbonate concretions, cemented layers, beds of laminated silt, and a 2 cm layer of limestone.

At a still higher level above the Diyana Plain, protruding from the smooth southwest flank of the beautiful anticlinal ridge (Kozik Dagh) north of the village of Diyana, is a prominent bench of resistant limestone breccia at least 15 m thick resting on Cretaceous shale. The remnant projects 1.5 km out from the mountain and has an elevation at about 1000 m above sea level at its outer edge--420 m above the river at Jindian only 2 km away. Scars on the anticlinal ridge show the source of the limestone debris. Comparable breccia benches occur on the opposite (northeast) flank of the same anticlinal ridge, at Havdian on the opposite side of the Diyana Plain, and at many other localities on the limestone ridges of the fold zone. The relative antiquity of the forms is indicated not only by their height above the present drainages but also, in the case of the Havdian occurrence, by the existence of a Mousterian cave (Babkhal) in the breccia (BRAIDWOOD, HOWE, *et al.*, 1960, p. 29, 60).

Both the breccia remnants and the 250 m-high alluvial remnant opposite Ruwandiz are probably much older than the gravel terraces of the Berserini gorges, and must date from the early Pleistocene.

Ruwandiz Gorge and Khalan Valley

After crossing the southeast edge of the Diyana Plain the Ruwandiz River enters the famous Ruwandiz Gorge, and obliquely crosses a limestone anticline for about 15 km in a canyon as much as 1000 m deep. It then flows longitudially along the next synclinal valley (the Khalan Valley) for 10 km, and in this segment shows well-developed terraces as much as 100 m above the river. These terraces carry a cover of up to 15 m of gravel rich in quartzite cobbles, and are overlain by limestone rubble graded from the adjacent ridges.

It is not possible at present to relate the Ruwandiz terraces of this longitudinal segment to the terraces of the Berserini Gorge and thus to mountain glaciation. The Khalan terraces are primarily erosional rather than depositional features: the alluvial cover is relatively thin. It is quite possible that these terraces record intermittent uplift of the highland area and are not climatically controlled.

The Ruwandiz River in Khalan Valley meets the Greater Zab River, which follows the same synclinal valley from the opposite end. The Greater Zab then turns at a right angle to cross the next anticlinal ridge, Berat Dagh, via Bekhme Gorge, and emerges onto the rolling Assyrian piedmont. Thence it traverses low folds in Tertiary conglomerate and siltstone, and constructs numerous erosional terraces with gravel veneers. The terrace topography gradually disappears downstream as the terrain becomes gentler towards the Tigris River, about 125 km from Bekhme Gorge. The Zab terraces in the Assyrian piedmont may also be tectonic rather than climatic.

Summary and Conclusions

The gorges and valleys of the Ruwandiz River contain the best exposed and most accessible Pleistocene terraces and associated colluvial deposits of any area in the mountains of northeastern Iraq.

The oldest Pleistocene deposits of the area are small benches of limestone breccia on the flanks of some of the limestone ridges. They represent old slopes and valley floors formed when the drainage level was as much as 400 m above the present.

The terraces of the lower part of the Ruwandiz River and in the Greater Zab River below the junction are probably also middle or early Pleistocene, but they may be tectonic rather than glacial-climatic.

The gravel terraces and intercalated colluvial deposits of the Ruwandiz River in the Berserini Gorge and upstream in the metamorphic zone are generally 30-50 m above the river. They are believed to be outwash deposits associated with a Pleistocene glacial phase earlier than the last. The glacial drift itself may be represented by till beds in the terrace deposits near Galala at an elevation of 1000 m, by the encrusted debris in the dissected old floor of the Marana Valley tributary, and by the smoothed till veneer in the open head of the Ruwandiz Valley above Rayat.

The last main glacial phase of the Pleistocene is represented by small cirques on sheltered slopes at elevations as low as 1500 m and probably even 1400 m and by larger cirques at 2100-3000 m. Fresh hummocky moraines are generally confined to the area just below cirques, but were found at elevations as low as 1200 m near Rayat. On the basis of topographic expression and elevation range, this glaciation is corelated with the last glacial phase.

Very high small upper-story cirques at 3000 m on the high ridge south of Marana Valley are assigned to very recent glacial erosion.

Cilo Dagh area

The Cilo Dagh (Jelo Dagh) and Sat Dagh in southeastern Turkey comprise the very highest sector of the Taurus-Zagros mountain arc in Kurdistan. The crests here reach 3800-4100 m above sea level but form neither the Turkish-Iraq political boundary nor the drainage divide. The Iraq frontier extends along the fold belt a few km to the south; the two main headwater tributaries of the Greater Zab River have their sources in the Anatolian Plateau to the north and cut through the metamorphic zone to the longitudinal valley that the Greater Zab follows as far as the junction with the Ruwandiz River in the Khalan Valley previously described.

Present and Pleistocene glaciers of the Cilo Dagh and Sat Dagh were studied by BOBEK (1940). He calculated the elevation of the present snowline in the area from 20 small glaciers which persist in the sheltered high valleys. The measurement was made in two ways: (a) average elevation between the terminus of the glacier and the crest of the ridge above the cirque, and (b) elevation of the lowest snow patches at the end of the melt season of observation (1937). The two types of measurement generally agreed and gave the following averages for the Cilo Dagh.

North slope, excessive shade	3100 m
North slope, average shade	3300 m
East slope and west slope	3400 m

Bobek also mapped the distribution of Pleistocene glaciers in these two ranges. The longest glacier was 10 km; one of the ice tongues reached an elevation of 1800 m. For the determination of Pleistocene snowline, he used the first of the two methods mentioned above--the average elevation between the Pleistocene moraines and the cirque ridges--and obtained the following averages for the Cilo Dagh:

North and northeast slopes	2600 m
West slopes	2500-2700 m
South slopes	2900-3000 m

From these figures an average of 700 m was presented by Bobek for the depression of the snowline during the Pleistocene. He compared this figure with the 1200 m usually assumed for the Alps, and suggested that the difference was the result of a drier climate. In fact he used this relation as evidence that there was no increase in precipitation in the Pleistocene in Iran.

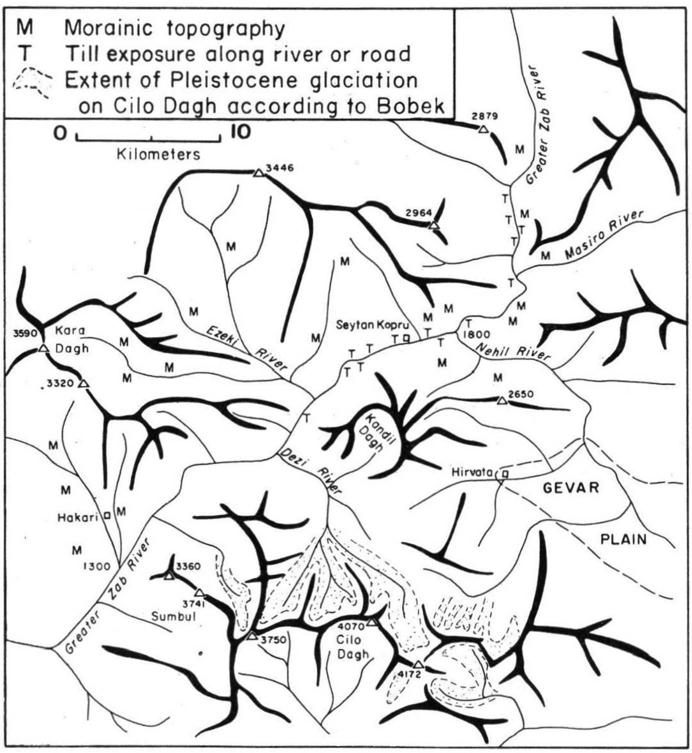

Fig. 13. Map of Cilo Dagh area, Taurus Mountains, Turkey, showing distribution of Pleistocene glacial features.

The writer's explorations in this region of the eastern Taurus of Turkey were confined to the region north and northwest of the Cilo Dagh (Fig. 13). The active glaciers and Pleistocene features of the high Cilo Dagh as mapped by Bobek where not visited, but Pleistocene glacial features were found over a broad area in and adjacent to the Greater Zab River valley at much lower elevations than those recorded by Bobek.

A 2700-3000 m ridge north of and parallel to the high Cilo Dagh bears a blanket of moraine over its entire north flank. This range (called Kandil Dagh on Bobek's map) is all above the tree line and bears a lush cover of tall grasses and herbs in the spring. Small bedrock knobs and short ridges project above the debris-covered slopes; at least one ridge is large enough to have supported a small cirque glacier at about 2300 m. Several small lakes and meadows on the slope indicate the youthfulness of the moraines. The glaciers that deposited the till presumably originated in cirques (not seen) near the crest of the range, and flowed down the slope in several lobes separated by visible medial moraines (Fig. 14). The ice filled the Nehil River valley at the north base of the Kandil Dagh, and

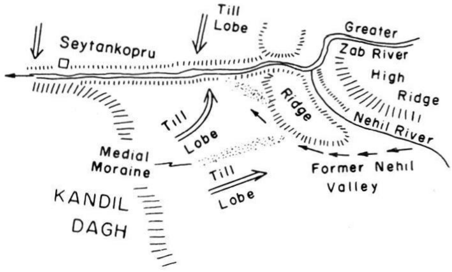

Fig. 14. Moraines in Greater Zab valley north of Kandil Dagh, Turkey. Diversion of Nehil River.

caused that river to be displaced northward and superposed onto the flank of the bedrock ridge to the north, where it has since cut a sharp gorge. The moraine extends down the old Nehil valley to the Zab River, where it is exposed at an elevation of about 1800 m east of Seytankopru.

Similar moraines can be observed on the flanks of other ridges at comparable heights north of Kandil Dagh toward the inner edge of the Taurus Range (see areas labeled M on Fig. 13). Small cirques can be spotted as low as 2300 m in this area. Till is exposed at many places along the Zab River, and it is clear that alternate tributary glaciers pushed the river to opposite sides of the gorge.

One such small tributary glacier, 2 km south of Ezeki River on the right bank of the Zab, headed in a sheltered east-facing cirque at about 2100 m (Fig. 15). A narrow gorge below the lip of the cirque leads down to a narrow drift-filled valley with transverse moraine loops and a lateral moraine 30 m high. The moraine loops terminate at an elevation of about 1700 m; beyond this the morainic fill descends rapidly for about 300 m

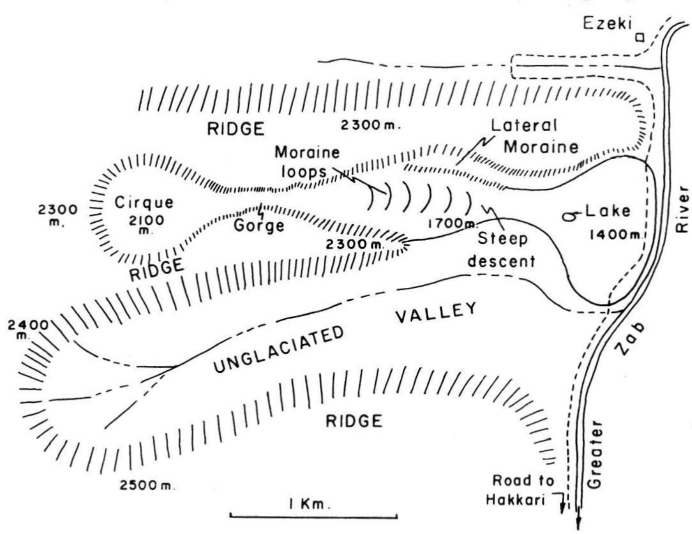

Fig. 15. Sketch of small glacial valley near Ezeki, ca. 16 km northeast of Hakkari, Turkey, showing moraine lobe in Greater Zab River.

and fans out in a broad lobe, crowding the Zab to the opposite side of its valley. A small lake on the middle of the lobe indicates the freshness and youth of the features.

Another interesting example of the effects of small tributary glaciers is seen near the village of Seytankopru, about 3 km down the Zab from the entrance of the Nehil River. At this point a double footbridge takes the main trail across the Zab to the villages on the north flank of the Kandil Dagh and thence to the Gevar Plain. A tributary glacier from the right side of the Zab plugged the valley with moraine and pushed the Zab against its left wall, causing a sharp gorge to be cut between the till and the quartz-schist bedrock of the left wall (Fig. 16). It so happened that at this point the nose of the moraine contained a huge block of green schist 15 m across. Continued erosion in the gorge beside the block

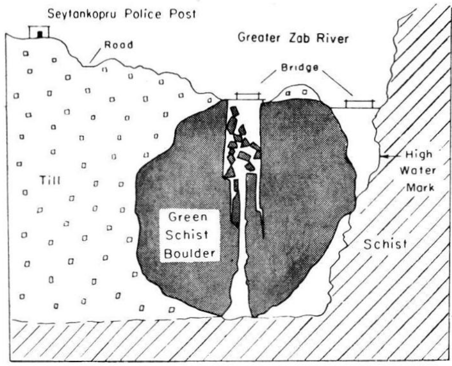

Fig. 16. Sketch across Greater Zab River near Seytankopru (ca. 26 km northeast of Hakkari, Turkey), showing gorge formed beside and within a huge boulder of schist at the nose of a moraine. Bridge is about 15 m above the river bottom. View upstream.

caused the block to be undermined. A fracture developed in the block, and the eastern half of it slumped over the little gorge, creating a bridge under which the stream now normally flows. The washing out of boulders from the till has produced a very steep rapids just below the bridge, resulting in a considerable congestion during highwater flow. At these times the water level rises about 15 m to fill completely the opening under the natural bridge and overflow the top. At this level the river is sufficiently high to occupy also the western channel that was produced when the part of the huge block had slumped. This western crack in the block shows a partial filling of boulders brought by the stream during flood time. The western overflow cut a broad arclike erosion scar at high level in the till bank just downstream from the block, and the oversteepening of the bank has resulted in a secondary slump scar farther to the west.

Farther downstream along the Zab where the dissection is deeper and the relief much greater, only a few tributary moraines reach the Zab River. The valleys heading in the Kara Dagh and other high ridges west of the Zab in this region appear from the distance to be greatly widened but choked with moraines (Fig. 17). The only such valley visited was that in which the town of Hakkari, the province capital, is located. This broad valley floor has an irregular topography and a cover of blocky debris, and the bounding ridges have many cirques. The Hakkari glacier extended probably about to the junction with the Zab, which has an estimated elevation of about 1500 m above sea level this area. The length of this glacier was about 10 km.

The Pleistocene snowline in the Cilo Dagh area is placed by the writer at 2100 m on the basis of the occurrence of cirques as low as this elevation. Some confirmation comes from calculations based on the mean elevation between end moraines and cirque-ridge, but this method is not considered so reliable because of the irregular crest line of the rugged mountain ridges and the gross inaccuracies of the available topographic maps for

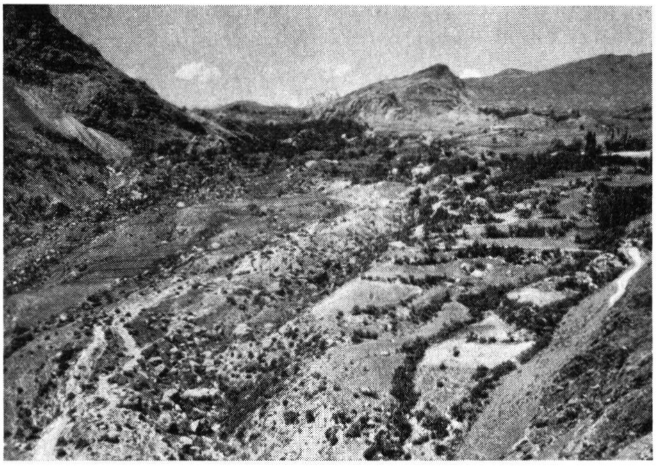

Fig. 17. Hakkari Valley, Taurus Mountains Turkey. Moraine-filled tributary of Greater Zab River, elevation ca. 1700 m. See Figure 13 for location.

the crestal area. If BOBEK's figure of 3300 m for modern snowline on shaded exposures is accepted, then the Pleistocene depression of snowline amounted to 1200 m rather than 700 m as calculated by BOBEK. The discrepancy in observation may result from the fact that BOBEK, having been attached to a mountaineering expedition, spent most of his time in the high mountains and did not have occasion to observe the glacial features in the lower hills and valleys enroute.

The new figure for Pleistocene snowline depression in the Cilo Dagh area (1200 m) is less than that for the Algurd Dagh area in adjacent Iraq (1800 m), described above, perhaps because of less precipitation in the northern shadow of the high mountains. The shadow effect today is easily seen in the vegetation of this region, and the Anatolian plateau is increasingly dry to the north. Many ridges rising to 3000-3300 m elevation in the plateau area between the Cilo Dagh and Lake Van show no signs of Pleistocene glaciation.

Other areas in the Taurus-Zagros Range

In the course of his explorations in Persia, the French archeologist DEMORGAN interpreted certain features as glacial in origin. In his voluminous reports, however, these features are mentioned only in footnotes (DEMORGAN 1909, p. 92) which refer to photographs in another memoir (DEMORGAN 1895). Exact locations and elevations are difficult to determine from these relations because of his use of obscure place names not identifiable on any maps available to the writer, who was aided in a careful search of place names by John J. ANDERSON.

The moraines pictured by DEMORGAN (1895, pl. 67, 80) in the Saidmarreh Valley are in fact the debris of a tremendous landslide that traveled 9 miles from its source area (HARRISON & FALCON 1937). Lake Gahar, east of Dorud at the south base of Kuh-i-Oshtoran (DEMORGAN 1895, pl. 71, 77; see also DESIO 1934) is not dammed by a moraine but by a landslide (FALCON 1946). Most of the other pertinent features mentioned by DEMORGAN are referred to as "glacial alluvium" or "glacial terraces"; those that could be located on topographic maps are indeed terraces, but a glacial source is by no means assured, and in fact seems unlikely.

The only certain locality of Pleistocene glaciation which the writer was able to find in southwestern Iran during his own field work in on the north-facing side of the isolated

Kuh-i-Oshtoran, a portion of the range termed Zardeh Kuh. The summit has an elevation of about 4200 m and the cirques have elevations estimated at about 3000 m. Moraines can be located in small valleys on the north slope down to about 2600 m, and there is a very large outwash fan at the north base of the mountain that is terraced by the modern river. These observations agree in part with those of DESIO (1934). FALCON (1946), however, denies the existence of true Pleistocene glacial deposits in the Zardeh Kuh.

Pleistocene snowline was apparently relatively high in the Zardeh Kuh because this region is farther inland from the Mediterranean source of moisture and because it is an isolated high mountain that does not attract so much moisture as a great mass of high ridges like the Cilo Dagh-Algurd Dagh area.

At the other end of the Taurus-Zagros Range, in south-central Turkey, Pleistocene glaciation has been studied by LOUIS (1944) as a part of a survey of the entire Anatolian Plateau and bounding mountain ranges. LOUIS adopted the figures of BOBEK for the eastern Taurus and prepared a contour map showing that Pleistocene snowline was below 2400 m around the margins of Turkey but rose in the interior to more than 2900 m at Erciyas Dagh in central Anatolia and more than 3200 m in the Lake Van-Lake Urmia region near the Iranian border.

Discussion

The new observations herein reported for the Kurdish mountains particularly in Iraq and Turkey require revision in the Pleistocene climatic reconstructions previously accepted. Snowline in the Cilo Dagh area was depressed in the Pleistocene approximately 1200 m to an elevation of 2100 m, and the glaciers extended down at least as low as 1500 m. In the Algurd Dagh-Ruwandiz area, snowline may have been depressed even more (1800 m), inasmuch as Pleistocene cirques were found at elevations as low as 1500 m above sea level, with till as low as 1100 m (Sideke). Even though the figures may be inaccurate because of poor maps and few observations, the depression of the Pleistocene snowline was much greater than that indicated by BOBEK (700 m).

The critical glacial features under discussion in the Kurdish mountains are assigned to the last glacial phase. To be conservative one might consider that some of the lowest cirques and till deposits should be referred to an earlier glaciation, but the figures for snowline would probably not greatly differ. Analogy may made with the Alps, where the Riss moraines on the northern foreland stand only a few km in front of the Würm, and are even locally overridden. Riss snowline depression is placed in the Alps only 100-200 m lower than the Würm (KLEBELSBERG, 1949, p. 683). Lack of deep weathering or of significant erosional modification of the low features in question in the Kurdish mountains suggest that they probably should be assigned to the last main glacial phase (Würm). Only the cemented and eroded terraces of the Ruwandiz River seem to be older.

Depression of the snowline is generally attributed to a reduction in atmospheric temperature, or to an increase in snowfall. Conversion of figures for Pleistocene snowline depression into estimates of temperature depression or changes in altitude distribution of life zones presents many problems. Much consideration has been given these matters in Europe over the years because so much information is available concerning Pleistocene features (WRIGHT, 1961). The present snowline in the Alps ranges from 2600 m on the north flank to 3100 m in the crestal region to 2800 m on the south flank (KLEBELSBERG, 1949, p. 662, Fig. 81). Pleistocene depression of the snowline is put at about 1200 m. If a vertical atmosperic temperature gradient (lapse rate) of 0.5° C/100 m is assumed, then the calculated Pleistocene temperature depression amounts to 6° C. On the other hand, in central Europe the changes in the distribution of certain plants and animals and the occurrence of certain frost features imply temperature depression of about 12° C. It has been suggested that the apparent discrepancy between 6° and 12° reflects the fact that the observations are made at different elevations above sea level, and that the lapse rate at low elevations is not so steep as it is in the vicinity of the snowline - - in fact may be isothermal or even seasonally inverted (MORTENSON, 1952; WRIGHT, 1961).

For the Kurdish montains, BOBEK inferred a 4° C temperature depression on the basis of a 700 m snowline depression. If the snowline was lowered as much as 1800 m instead of 700 m, a temperature depression of 9° C is implied if the usual gradient of 0.5° C/ 100 m is assumed. As a matter of fact, this gradient may not be correct for semi-arid regions. For ground stations in the semi-arid mountains of western United States, BAKER (1944, p. 225) calculated a mean annual lapse rate of 0.6° C/100 m, and a mean July rate of 0.65° C/100 m. For the Zagros Mountbains the data are sparse, but a series of 20 short-period stations in southwestern Iran in a transect from the Mesopotamian piedmont across the Zagros Range to the Iranian Plateau shows mean annual and mean July lapse rates of about 0.75° C/100 m (Table 1, Fig. 3). Inasmuch as there is a 2-degree latitudinal pread between piedmont and plateau stations, a latitudinal correction of 0.7° C per degree Latitude might be applied to the curves, yielding a corrected figure of about 0.67° C/100 m. Use of such a gradient with the 1800 m of snowline depression implies a temperature depression of a least 12° C.

Even if Pleistocene temperature may have deeply depressed in the mountains in Kurdistan, it is difficult to believe that the mean annual temperature in the Kurdish piedmont and the Mesopotamian lowland was 12° C lower than the present. No positive geologic or paleontologic evidence can be adduced to support such a change. Frost features have been utilized elsewhere as temperature indicators, and TROLL (1947, Fig. 1) has shown that the lower limit of modern frost soils parallels the snowline at slightly lower elevations. The distribution and character of certain frost soils (especially involutions and fossil ice-wedges) have been used as the basis for a paleo-climatic map of Europe (POSER, 1948; FRENZEL, 1958; WRIGHT, 1961). In the Kurdish mountains the only features studied that might be attributed to intensified frost action and solifluction on unstable slopes above the treeline are the thick colluvial deposits in the Ruwandiz valley.

MORTENSEN (1957) has suggested that the lapse rate for lower elevations in subtropical latitudes is steeper than at elevations close to the snowline, so that a large temperature depression near the snowline was rapidly reduced to a small depression in the lowlands - - the opposite effect to that he postulated for periglacial Europe where ground cooling was critical (MORTENSEN 1952). For the Kurdish mountains there are too few temperature stations to determine whether the rate is greater in the lowlands than in the mountains: Figure 3, however, gives no hint such a trend. No free-air lapse rates extending to higher elevations are available for this region. MORTENSON's hypothesis therefore cannot be tested further with the data at hand.

The effect of change in snowfall on the elevation of snowline is difficult to segregate from the effect of change of temperature. An attempt was made by KLEIN (1953) for Europe, but the figures on which it was based are subject to many errors (WRIGHT, 1961). Snowline is clearly higher in regions of low snowfall. KLUTE (1928) prepared meridional profiles through mountain groups in the eastern and western hemispheres to show the relationship of present and Pleistocene snowline to latitude and precipitation. These curves show that in general the snowline is low in the polar regions because it is cold, moderately low in the middle latitudes because of the greater precipitation in the zone of westerly winds, high in the subtropical arid belt, and not quite so high in the wet tropical mountains. In explaining the relation between snowline and snowfall, KLUTE and also PASCHINGER (1923) concluded that the relatively great depression of snowline in middle latitudes was a result of the fact that in high mountain ranges in humid regions the zone of maximum precipitation is at moderate elevations, especially in winter, because the invading air masses are not forced to rise very high before losing most of their moisture. In this view, in the Pleistocene a modest temperature depression caused the snowline to be lowered down into the zone of maximum precipitation, where the additional snow supply resulted in further lowering of the snowline. KLEBELSBERG (1949, p. 664), however, denies that such a zone of maximum snowfall exists at intermediate elevations in the Alps. In the northern Sierra Nevada of California, early work suggested that precipitation increases at an average rate of 75 mm/100 m rise in elevation up to an elevation 2000 m above sea level, above which it decreases (PALMER, 1915). It has recently been claimed, however, that the snow gauges by which this gradient was determined were inaccurate, and that actually the snowfall increases in amount all the way to the crest (MILLER, 1955, p. 22).

For the Kurdish mountains, conditions of greater snowfall must be postulated if the inferred 1200-1800 m depression of snowfall was not a result entirely of temperature depression. It is certain that the Mediterranean Sea was the source of moisture in the Pleistocene just as today, for the Pleistocene just as today, for the Pleistocene snowline was much lower on the south side of the range than on the interior plateau north of the range. In fact, BOBEK (1937, p. 178; 1954, p. 18) believes that because of the strengthened Siberian anticyclone during the last glaciation the Iranian Plateau was cold but no moister than today, and thus that the term "pluvial" should not be applied to the Pleistocene climate of this region. The stronger temperature gradient between the Iranian Plateau and the Mesopotamian Lowland may have intensified the storminess on the outer flank of the intervening mountains. These mountains are far enough inland so that the low winter temperatures characteristic of a continental climate could prevail, but there was enough orographic lifting to provide heavy snowfall.

LOUIS's (1944, p. 476) map of Pleistocene snowline for Turkey clearly shows the effects of precipitation in that region. The contours are concentric around the dry interior part of the country. LOUIS determined that Pleistocene depression of the snowline ranges from 1000 m on the more humid outer margin of the area to 700 m in the continental interior. He pointed out (1944, p. 477) that steep changes in values for snowline depression are found in areas where the high mountains face the prevailing storms from the west. Thus through most of the western and central Taurus as well as in the ranges of northern Anatolia the grain of the topography trends parallel to the storm tracks, and the precipitation contrast toward the inland is not great. But in the eastern Taurus and the Zagros ranges the trend bends to the southeast. Perhaps the new figures for snowline depression in this area may be explained in part by this relation. The moutain axis is not only favorably oriented to receive more moisture from the Mediterranean storms but it is also higher and more massive. Furthermore, the more continental location farther from the sea may mean that the winter temperatures in the Pleistocene were lower than they were farther seaward; a greater proportion of the precipitation therefore may have fallen as snow, and the winter season may have been longer. According to BOESCH (1941) the snowfalls occur at present in association with the Mediterranean cyclones rather than with the Arabian anticyclones. The Pleistocene climate therefore may have involved a greater incidence or deeper penetration of Mediterranean storms. Such a situation checks with the evidence from Europe that that region was generally marked by high pressure in summer as well as winter (POSER, 1950), and that the major storm tracks were pushed south of the Alps. Intensified fall and spring storms would increase the snowpack in the Kurdish mountains, and occasional summer disturbances might introduce enough cloudiness to deter summer melting.

Farther southeast along the Zagros ranges the region becomes progressively drier, for several reasons. The mountains are generally lower, except for the Zardeh Kuh, which still bears perennial snow patches. The distance from the Mediterranean moisture source is greater. The temperatures are higher because of the more southerly latitude. For the Pleistocene no certain traces of glaciation have been found except in the Zardeh Kuh, where moraines occur as low as an estimated 2600 m on the north flank of these mountains east of Dorud.

The Pleistocene snowline depression in the Kurdish mountains (as much as 1800 m) differs from that in the Alps (1200 m) perhaps because of different precipitation regimes in the two areas and the effect of Pleistocene climatic changes on those regimes. The Alps are marked by a summer precipitation maximum, with only about 20% of the total precipitation occurring in the three winter months (Table 3), all presumably as snow at

Table 3

Temperature and precipitation for selected stations in the Sierra Nevada (California) and the Alps

	LAT.	ELEV.	TEMPERATURE °C				PRECIPITATION		
			Mean	Mean	Mean	Range	Annual	DJF	NDJFM
	°	m.	Ann.	Jan.	July		m	%	%
SIERRA NEVADA, California (1) Snowline 3500 m, treeline 3300 m									
Twin Lakes, Alpine Co.	39	2373	5	—4.4	14.6	19	1042	53	75
Cisco, Placer Co.	39	1733	9	0.4	18.4	18	1391	53	81
Mt. Shasta, Siskayou Co.	41	1074	9	0.6	18.7	18	870	49	76
Yosemite, Mariposa Co.	38	1208	10	1.8	21.2	19	861	52	76
Downieville, Sierra Co.	40	877	12	3.0	20.2	17	1543	56	80
ALPS (2) Snowline 2700 m, treeline 2000 m									
Zürich	47	477	8.7	— 0.3	18.0	18	1106	17	29
Obir	47	2044	0.4	— 6.9	9.0	16	1530	16	31
Säntis	47	2500	—2.3	— 8.6	4.9	14	2720	23	36
Sonnblick	47	3106	—6.3	—12.8	1.0	13	1572	24	39

(1) Data from U. S. Dept. Agric., 1941 NDJFM = November through March

(2) Data from CLAYTON (1944—1947) DJF = December, January, February

elevations appropriate for glaciation. In the Kurdish montains, however, 52% of the precipitation comes during the winter. If November and March are included with the winter, the average figure becomes 79%, compared to about 35% for the Alps for the same five months. Thus although the total precipitation in the Kurdish Mountains is less than in the Alps at comparable elevations, the total snowfall may be as great - and this is the iportant factor in glaciation. It is difficult to evaluate whether clear summer skies with bright sun (Kurdistan, 37° N. Lat.) is more effective or less effective in melting snow than summer rains (Alps, 47° N. Lat.), but at any rate the present snowline in the Kurdish mountains (3100-3400 m) is not much higher than that in the Alps (2600-3100 m) despite the difference in latitude.

An area more comparable to the Kurdish mountains in precipitation regime is the Sierra Nevada of California, which also has a Mediterranean climate with winter snow and summer drought. Here 53% of the precipitation comes in the months December, January, and February, and 78% if November and March are included (Table 3). The total snowfall is among the greatest in the United States, generally exceeding 10 m in thickness at an elevation of 2000 m, and sometimes reaching 18 m (PALMER, 1915; MATTHES, 1930, p. 10). Despite the great accumulation of snow, melting is rapid in spring and summer and even in winter because of the radiational heating from the coniferous forest (treeline is about 3200 m) and the persistent inflow of warm Pacific air (MILLER, 1955, p. 184). The pressent snowline, as represented by many small glaciers in sheltered cirques high in the range in the Yosemite Park area (38° N. Lat.), is at about 3500 m. The Pleistocene snowline, as recorded by the lowest small cirques visible on the topographic maps, is at about 2100 m in the same area. Pleistocene depression of snowline therefore amounted to about 1400 m. The effect of precipitation can be seen from the fact that the Pleistocene snowline rises from 1800 m in the north at 40° Latitude to more than 2700 m in the south at 36° Latitude. The 4° latitudinal difference is equivalent to a mean temperature difference of about 3° C, which would be equivalent to about 500 m of snowline differential if a vertical temperature gradient of about 0.6° C/100 m is assumed. The remaining 400 m of difference must be attributed to the difference in precipitation which today ranges from 1500 mm in the north to 1000 m in the south. A similar precipitation effect is seen in the comparison of the Pleistocene snowline at the northern end of the range (1800 m with 1500 mm precipitation) with that in the Ruby-East Humboldt Range in Nevada at the same latitude but 400 km to the east, where the Pleistocene snowline was 2800 m (SHARP 1938, p. 321) and the present precipitation only 550 mm.

Other mountain ranges with a Mediterranean-type precipitation regime include the High Atlas of northwest Africa (34° Lat), where BÜDEL (1952) puts the modern snowline at 3300 m. Pleistocene cirques occur down to 1900 m, so a depression of 1400 m is inferred for the Pleistocene. Elsewhere in low latitudes, recent work by PASCHINGER (1954) in the Sierra Nevada of Spain suggests a Pleistocene snowline depression of 1200-1400 m, HÖVERMANN (1954) 1700 m in Abyssinia, and HEUBERGER (1956) 2100 m in the Himalaya of East Nepal. In most of these regions the forest cover is open and at relatively low elevations. The arboreal radiation factor which promotes spring melting in the coniferous forest region of the Sierra Nevada in California (MILLER 1955, p. 89 ff.) therefore is not effective here.

Another basis extensively used for determination of Pleistocene climate is vegetational reconstruction through pollen analysis. BÜDEL (1951) has utilized such data, along with information on frost features and loess deposits, for a paleo-ecological map of Europe, and FRENZEL (1958) has done the same for northern Eurasia. Such data for the Kurdish mountains are lacking completely, so there is no evidence in this region to supplement the snowline data on Pleistocene climate. Pleistocene vegetational maps that have been extended to cover the eastern Mediterranean region show a broader forest cover in the Kurdish Mountains than the present and a broader shrub steppe cover in Mesopotamia (WISSMANN 1956; BUTZER 1958, p. 140). It should be emphasized that these are based on inference from local geologic features or on extrapolation from Europe, for there is no paleobotanical information.

Studies in Europe have shown that the entire area between the northern ice sheet and the Alps was essentially unforested and that on the south flank of the Alps the treeline was lowered even more than the snowline (FIRBAS, 1939, p. 105). In the Kurdish mountains it seems unlikely that the upper treeline, which now is located at an elevation of about 2000 m above sea level, was lowered as much as the snowline was lowered, for this would imply that the etmperatures even in Mesopotamia were too low for tree growth. The vertebrate fossils from both the Mousterian and Upper Paleolithic cave deposits in Kurdistan, however, suggest that the late Pleistocene fauna was essentially the same as the recent (REED & BRAIDWOOD 1960, p. 169);. Some of the animals, of course, undoubtedly were hunted in different life zones from those in which the caves occur, but there seems to be a direct conflict between this line of evidence and that supplied by snowline depression. A possible explanation rests in the suggestion of SOLECKI (1955, and personal conversation, 1960) that at the cave near Shanidar in Kurdistan (see Fig. 2) an unconformity between the Upper Paleolithic (C-14 dated at 29,000 years ago) and the Mesolithic (12,000 years ago) implies that the cave was uninhabited during this time interval, perhaps because the treeline had descended below this elevation (700 m). If correlation be made with the Alps, this was the time of maximum glaciation and life-zone displacement in the mountains. We must await some more direct paleobotanical studies (*e. g.*, pollen analysis) before further attempts to determine the Pleistocene altitude limits of the vegetation zones.

It also should be emphasized that the Kurdish mountains are located in any area of strong climatic gradient today and probably also in the Pleistocene. The contrast between the relatively well-watered outer flank of the Kurdish mountains and the dry and continental Iranian Plateau bordering the inner flank is very great, as can be seen from the modern climatic data and vegetation and from the steep inland rise of the Pleistocene snowline. BOBEK (1937, 1954) has made a case for a Pleistocene climate on the Plateau that was colder (for advance of glaciers) but not more humid *(i. e.,* not a pluvial climate), and explains expansion of inland lakes solely on the basis of temperature change. We must therefore set this conclusion against the new evidence from the outer flank of the Kurdish mountains for snowline depression that can only be accounted for by increased precipitation as well as by lower temperatures. It is possible that both these conclusions are correct - - if we accept the proposal that we are dealing with two separate climatic provinces whose contrasts were accentuated during the Pleistocene. The increased frequency and intensity of the cyclonic disturbances that entered Mesopotamia in the Pleistocene could account for increased snowfall on the outer flank of the Kurdish mountains, but the intensified Siberian anticyclone in winter could block the penetration of these storms into the Iranian Plateau.

The effects of late Pleistocene climatic changes on prehistoric man in the Kurdish mountains and piedmont are even more problematical than the effects of the vegetation - - especially the problem of the origin of the village-farming community. A critical factor is the timing.

The absolute dating of Pleistocene and Recent glacial and early archeological materials is best accomplished by the radiocarbon method. Unfortunately, no organic materials suitable for dating were found in any of the glacial deposits described herein. Dates from the Paleolithic cave of Shanidar in the Baradost Dagh on the Greater Zab River (Fig. 5) may be related to an internal stratigraphy of stalagmite and other climatically controlled deposits (SOLECKI 1955), but these in turn have not been correlated with the glacial features in the mountains. At present they serve best to indicate the time range of the Paleolithic cultures in this region.

Lacking radiocarbon dating of the glacial deposits in the mountains, one must resort to correlation of the features with those of the nearest comparable area where a chronology is available, namely the Alps. The glacial features in the Kurdish mountains, especially where observed in the Algurd Dagh area and north of the Cilo Dagh, compare to the deposits of the Würm (last) glaciation in the Alps with respect to degree of weathering and erosional modification. Discrete successive moraine loops such as are visible near the termini of the Alpine Würm glaciers could not be identified in the Kurdish mountains, but this is not surprising in view of the fact that the glacier fronts were still well up in the mountains and that the bedrock topography was irregular. No attempt could be made to subdivide the last glacial phase in the Kurdish mountains - - certainly no attempt at correlation with subdivisions of the Würm stage within the Alps. The significance of the very fresh cirques high on the older headwalls has not been fully evaluated.

The Alpine glaciers probably reached their Würm maximum about 20,000 years ago (GROSS 1958, p. 173) - - this dating unfortunately is not based on C-14 dates of the Alpine deposits themselves but instead is based on correlation with carbon-dated last-glacial deposits of northern Germany and central United States. Retreatal positios of the Alpine ice fronts during the succeeding millenia are correlated with less accuracy, but at the time of the "Schlussvereisung", believed to be correlated with the Central Swedish moraine of the north and the "Younger Tundra" zone of the pollen sequence about 11,000 years ago just after the Alleröd oscillation, the Alpine ice sheet had disintegrated and local glaciers reformed. KLEBELSBERG (1949, p. 706) places the snowline for this time at 800-900 m below the present, compared to 1200 m below the present for the Würm maximum, and

Zagwijn (1952, p. 23) suggests on the basis of a pollen study near Innsbruck that the treeline was more than 1000 m lower than the present. These fiugures imply that the climate was still very cool in the Alpine foothills. The more abundant pollen-analytical evidence farther north in Europe, however, indicates that during the Alleröd interval about 12,000 years ago the July temperature in Denmark was only about 3° C less than today (Iversen, 1954 p. 98), compared to a depression of 12° C for the Würm maximum. A similar figure is given by Firbas (1949, p. 287) for southern Germany. For the Younger Tundra phase that followed, these authors suggest July temperatures about 6° C below the present. Attainment of postglacial temperature at least as warm as today was achieved perhaps 9000 years ago.

If the Alps are to be used as a standard of comparison for the late-glacial of Kurdistan, then, we must conclude that the climate had already improved greatly by 11,000 years ago, and that by 9000 years ago the cold period had passed completely. The Kurdish glaciers should be even more sensitive to climatic changes than the Alpine because they were smaller - - the Alps had a fairly large and continuous ice cap that may have accentuated the severity of the local climate.

Radiocarbon dates of Upper Paleolithic and later prehistoric cultures in the Kurdish mountains and piedmont suggest that 12,000 years ago man was still essentially a cave-dweller and hunter in this region, but that by 11,000 years ago he was occupying open sites well back in the mountains (*e. g.*, Zawi Chemi near Shanidar) as well as in the piedmont (Karim Shahir near Jarmo in the foothills above Kirkuk) (Braidwood & Howe 1960, p. 157). This is the time of incipient cultivation, followed in the next two or three millenia by the development of the village-farming community such as Jarmo.

Bobek (1954) has considered this problem for the Iranian plateau, premised on the belief that the early post-glacial (ca. 11,000 to 6,000 years ago) was marked by a dry climate. The principal evidence for a dry climate for this time is the occurrence of loess deposits near the northern base of the Elburz Mountains at elevations below the late Pleistocene shoreline of the Caspian Sea (Bobek 1937, p. 174 ff). The loess deposits, which contain "Neolithic" or Copper-age artifacts, are believed to require steppe conditions for formation, yet the deposits have since been invaded by forest. These relations by themselves are convincing, and Bobek attempts to support them with other paleo-climatic inferences from the Iranian Plateau proper. Although the other features are impressive - *e. g.*, sand dunes resting on lake beds - none of them has been dated, and his case for a dry period during the time of inception of agriculture must rest largely on the Caspian loess relations.

Bobek's conclusions for a pronounced post-glacial dry period in Iran are not necessarily applicable to the outer flank of the Kurdish mountains, which is in a basically different climatic province. He notes that most of the early agricultural village sites in the Mesopotamian piedmont are located in the present steppe or forest steppe where the precipitation today is estimated to be greater by 100-150 mm than the minimum that is required for dry farming (250-300 mm), implying that the rainfall has increased by this amount since the sites were occupied. Too many assumptions are necessary for this observation to be accepted as positive evidence for climatic change, however.

Butzer (1958, p. 103-128) has put together for the entire Near East the scattered geologic and archeologic suggestions of past climates. He has constructed a chronology involving several shifts of wet and dry during postglacial time. The evidence is so scattered geographically and climatically (from the North African coast to the Iranian Plateau), so difficult to place in time, and in cases so contradictory that one hesitates to accept such chronology as a regional framework that can be applied to such a tenuous anthropological matter as domestication of plants and animals and the beginnings of village life. When so little is known about the detailed distribution and sequence of the cultures themselves, it may be unsafe to correlate the cultures with a climatic chronology that is equally unsure.

The writer has developed the general impression that in the outer foothills of the Kurdish mountains and in the Mesopotamian piedmont the geologic record of post-glacial climatic changes is very poor - - probably because the climatic changes were relatively weak. Most streams have not incised very deeply into their Pleistocene terraces and many prehistoric mounds located on low terraces only a few meters above the modern floodplains have not been flooded since their occupation 5000-8000 years ago. Mounds exposed and unoccupied since their accumulation as much as 10,000 years ago show a soil on the surface, but otherwise little modification. Nor do they bear a mantle of loess despite the fact that long-lasting dust storms are characteristic of present-day spring weather in this area.

In only one small stream system in the Kurdish foothills has the writer been able to find datable geologic features that imply post-glacial climatic change (WRIGHT 1952). This area, the Chemchemal valley northeast of Kirkuk, contains an impressive cut-and-fill terrace sequence in which the main fill surface is dated as late Pleistocene (post-Acheulian, pre-Jarmo) and a younger fill terrace is distinctly post-glacial (Assyrian). The Assyrian cut and fill terrace resembles the arroyos of the American Southwest, which are associated with archeological sites and are attributed to the effect of minor climatic changes on soil erosion and stream regime during the last 2000 years (BRYAN 1941).

Valleys with such thick unconsolidated sediment located in the proper portion of a drainage basin and in a suitable vegetation zone may be particularly susceptible to changing hydrologic conditions and thus may record postglacial climatic changes, but the paucity of such features in the Kurdish piedmont and foothills suggests either that requisite conditions did not prevail or that the climatic changes were not sufficiently pronounced. As striking as are the erosional cycles in the American Southwest, the events are not well represented otherwise in the paleoclimatic record, and the last cycle of erosion, which is historically dated as beginning 1880-1900, is so poorly recorded in the weather data of nearby stations that a controversy has long existed whether this erosional cycle is a result of climatic change or of overgrazing by sheep (see discussion in ANTEVS 1952, and PETERSON 1950). Although the earlier erosional cycles in the Southwest may have been accompanied by abandonment of villages (BRYAN 1941) and thus movements of peoples, the habitation in this region (which is a desert steppe much drier then the Mesopotamian area under consideration) was hazardous at best.

In conclusion, the writer believes that the evidence from glaciation in the Kurdish mountains indicates that the Pleistocene climate was probably colder and wetter in the foothills and piedmont , but that the change to a postglacial climate much like the present was probably essentially complete by the time of the beginnings of cultivation and the establishment of permanent villages 11,000-9,000 years ago. Even in the late Pleistocene before the complete recession of the mountain glaciers there must have been ecological environments just as suitable for this new economy as there were a few thousand years later, because the effect of climatic change was primarily to raise or lower the various life zones in the foothills and mountains, and early man should be able to follow the zone to which he was most adapted at the particular stage of his cultural development. The same reasoning applies with greater force to the post-Pleistocene, for here we have less evidence for climatic change and the related shifting of the life zones, and at the same time we know that after the great agricultural revolution the pace of cultural development was accelerated. By this time, there were many sociological factors controlling cultural change, such as increasing abilities to control the environment through irrigation, transportation, etc. The physical and ecological factors thus may have been largely submerged. Certainly in marginal living areas where the economy was at a subsistence level a series of bad years (perhaps representing a minor climatic fluctuation) might bring about disasters or even impel changes in the economy, but it could as easily encourage migration to more favorable regions. In the earlier periods before the area was broadly populated - - perhaps in the times before mounted nomadism became a common way of life (BOBEK) 1954, p. 18 ff) - - there were many areas suitable for habitation without competition from other peoples. It may only have been in recent millenia that population pressures have been sufficiently high to restrict the migration of cultural groups. Although some such migrations or invasions may have been stimulated by local climatic changes (BROOKS 1949, p. 281 ff) many others must be attributed to sociological or political motivations.

References cited

ANTEVS, Ernst (1952): Arroyo-cutting and filling. - Journ. Geology **60**, p. 375-385.

BAKER, F. S. (1944): Mountain climates of the western United States. - Ecological Monographs. **14**, p. 229-243.

BOBEK, Hans (1937): Die Rolle der Eiszeit in Nordwestiran. - Z. f. Gletscherk. **25**, S. 130-183. - - (1940): Die gegenwärtige und eiszeitliche Vergletscherung im Zentralkurdischen Hochgebirge. - Z. f. Gletscherkunde **27**, S. 50-88. - - (1951): Die natürlichen Wälder und Gehölzfluren Irans. - Bonner Geogr. Abhandl. **8**, 62 S. - - (1952): Beiträge zur klimaökologischen Gliederung Irans. - Erdkunde **6**, S. 65-84. - - (1954): Klima und Landschaft Irans in vor- und frühgeschichtlicher Zeit. - Geogr. Jahresber. Österr. **25**, S. 1-42.

BOESCH, H. H. (1941): Das Klima des Nahen Ostens. - Vierteljahresschr. d. Naturforsch. Ges. Zürich **86**, S. 8-61.

BRAIDWOOD, R. J., HOWE, BRUCE *et al* (1960): Prehistoric investigations in Iraqi Kurdistan. - Oriental Institute of University of Chicago, Studies in Ancient Oriental Civilization No. 31, 184 p.

BROOKS, C. E. P. (1949): Climate through the ages. - London: Ernest Benn, 2nd ed., 395 p.

BRYAN, Kirk (1941): Pre-Columbian agriculture in the Southwest, as conditioned by periods of alluviation. - Assoc. Amer. Geographers Annals **31**, p. 219-242.

BÜDEL, Julius (1951): Die Klimazonen des Eiszeitalters. - Eiszeitalter und Gegenwart **1**, S. 16-26. - - (1952): Bericht über klima-morphologische und Eiszeitforschungen in Niederafrika. - Erdkunde **6**, S. 104-131.

BUTZER, K. W. (1958): Quaternary stratigraphy and climate in the Near East. - Bonner Geogr. Abhandl. **24**, 157 S.

CLAYTON, H. H. (1944-1947): World weather records. - Smithsonian Misc. Coll. **79** (1944), **90** (1944), and **105** (1947).

DEMORGAN, Jacques (1895): Mission scientifique en Perse. - Etudes Geographiques **2**, 331 p., Paris: Ernest Leroux. - - (1909): Les Premiers civilizations. - Paris: Ernest Leroux, 508 p.

DENNIS, P. E. (1953): Report to the Government of Iraq on the investigation and development of ground-water resources. - Food and Agriculture Organization of the United Nations, Rome, 49 p.

DESIO, A. (1934): Appunti geografici e geologici sulla catena dello Zardeh Kuh in Persia. - Mem. Geol. Geogr. di Giotto Dainelli **4**, Firenzi.

EL FANDY, M. G. (1946): Barometric lows of Cyprus. - Royal Met. Soc., Quart. Journ. **72**, p. 291-306.

FALCON, N. L. (1946): The evidence for a former glaciation in the south-west Persian mountain belt. - Geogr. Journ. **107**, p. 78-79.

FIRBAS, Franz (1939): Vegetationsentwicklung und Klimawechsel in der Mitteleuropäischen Spät- und Nacheiszeit. - Naturwissenschaften **27**, S. 81-89, 104-108. - - (1949): Waldgeschichte Mitteleuropas. - Jena: Gustav Fischer, **1**, 480 S.

FISH, W. D. & DUBERTRET, Louis (1944): Carte pluviometrique du Moyen Orient. - Del. Gen. de France du Levant, Sect. Geol., Notes et Mem. **4**, p. 115-121.

FRENZEL, B. (1959): Die Vegetations- und Landschaftszonen Nord-Eurasiens während der letzten Eiszeit und während der postglazialen Wärmezeit. - I. Teil: Allgemeine Grundlagen. Akad. Wiss. Lit. Mainz, Abh. Math.-Nat. Kl., Nr. 13, S. 935-1099.

GANJI, M. H. (1960): Iranian rainfall data. - Univ. Tehran Arid Zone Research Centre, Publ. 3, 191 p.

GROSS, Hugo (1958): Die bisherigen Ergebnisse von C^{14}-Messungen und paläolithischen Untersuchungen für die Gliederung und Chronologie des Jungpleistozäns in Mitteleuropa und den Nachbargebieten. - Eisz. und Gegenw. **9**, S. 155-187.

HARRISON, J. V. & FALCON, N. L. (1937): The Saidmarreh landslip, Southwest Iran. - Geogr. Journ. **89**, p. 42-47, London.

HEUBERGER, Helmut (1956): Beobachtungen über die heutige und eiszeitliche Vergletscherung in Ost-Nepal. - Z. f. Gletscherkunde u. Glazialgeologie **3**, S. 349-364.
HÖVERMANN, Jürgen (1954): Über die Höhenlage der Schneegrenze in Äthiopien und ihre Schwankungen in historischer Zeit. - Nachr. Akad. Wiss. Göttingen, Math.-Phys. Kl., Nr. **6**.
IVERSEN, Johs. (1954): The Late-glacial flora of Denmark and its relation to climate and soil. - Danmarks Geol. Unders., R. II, no. 80, p. 87-119.
KLEBELSBERG, Raimund von (1949): Handbuch der Gletscherkunde und Glazialgeologie. - Vienna: Springer Verlag. Bd. I, Historisch-regionaler Teil, S. 407-1028.
KLEIN, Annemarie (1953): Die Niederschläge in Europa im Maximum der Letzten Eiszeit. - Peterm. Geogr. Mitt. **97**, S. 98-104.
KLUTE, Fritz (1928): Die Bedeutung der Depression der Schneegrenze für eiszeitliche Probleme. - Z. f. Gletscherkunde **16**, S. 70-93.
LEMBKE, H. (1940): Eine neue Karte des Jahresniederschlages im westlichen Vorderasien. - Peterm. Geogr. Mitt. **86**, S. 217-225.
LOUIS, Herbert (1944): Die Spuren eiszeitlicher Vergletscherung in Anatolien. - Geol. Rundschau **34**, S. 447-481.
MATTHES, Francois (1930): Geologic history of the Yosemite Valley, California. - U. S. Geol. Survey, Prof. Paper 160, 137 p.
MILLER, D. H. (1955): Snow cover and climate in the Sierra Nevada, California. - Univ. Calif. Publ. Geogr. **11**, p. 1-218.
MORTENSEN, Hans (1952): Heutiger Firnrückgang und Eiszeitklima. - Erdkunde **6**, S. 145-160. - - (1957): Temperaturgradient und Eiszeitklima am Beispiel der pleistozänen Schneegrenzdepression in den Rand- und Subtropen. - Z. f. Geomorphologie **1**, S. 44-56.
PALMER, A. H. (1915): Region of greatest snowfall in the United States. - Monthly Weather Review **43**, p. 217-221, Washington.
PASCHINGER, Herbert (1954): Würmvereisung und Spätglazial in der Sierra Nevada (Spanien). - Z. f. Gletscherk. u. Glazialgeol. **3**, S. 55-67.
PASCHINGER, Viktor (1923): Die Eiszeit ein meteoroligscher Zyklus. - Z. f. Gletscherkunde **13**, S. 29-65.
PABOT, Henri (1960): Native vegetation and its ecology in the Khuzistan River basins. - Ahwaz, Iran: Khuzistan Development Service, unpublished report.
PETERSON, H. V. (1950): The problem of gullying in Western valleys. - Ch. 23, p. 407-436, in: Applied Sedimentation (P. Trask, ed.), New York: Wiley, 707 p.
POSER, Hans (1948): Boden- und Klimaverhältnisse in Mittel- und Westeuropa während der Würmeiszeit. - Erdkunde **2**, S. 53-68. - - (1950): Zur Rekonstruktion der spätglazialen Luftdruckverhältnisse in Mittel- und West-Europa auf Grund der vorzeitlichen Dünen. - Erdkunde **4**, S. 81-88.
REED, C. A. & BRAIDWOOD, R. J. (1960): Toward the reconstruction of the enviromental sequence of northeastern Iraq. - Oriental Inst. Univ. Chicago, Studies in Ancient Oriental Civilization. No. 31, p. 163-174.
SHARP, R. P. (1938): Pleistocene glaciation in the Ruby-East Humboldt Range, northeastern Nevada. - Journ. Geomorphology **1**, p. 296-323.
SOLECKI, Ralph (1955): Shanidar Cave, a Paleolithic site in northern Iraq. - Smithsonian Institution, Annual Report 1954, p. 389-425.
TROLL, K. (1947): Die Formen der Solifluktion und die periglaziale Bodenabtragung. - Erdkunde **1**, S. 162-175.
U. S. Dept. Agriculture (1941): Climate and Man. - U. S. Dept. Agriculture, Yearbook of Agriculture, 1248 p.
WISSMAN, Hermann von (1956): On the role of nature and man in changing the face of the dry belt of Asia. - In: Thomas, W. L. (editor), Man's role in changing the face of the earth, p. 278-303, Chicago: University of Chicago Press.
WRIGHT, H. E. (1952): Geologic setting of four prehistoric sites in northeastern Iraq. - Amer. Schools of Oriental Research, Bull. 128, p. 11-24. - - (1961): Late Pleistocene climate of Europe: a review. - Geol. Soc. America, Bull. **72**, p. 933-983.
ZAGWIJN, W. H. (1952): Pollenanalytische Untersuchung einer spätglazialen Seeablagerung aus Tirol. - Geologie en Mijnbouw **14**, p. 235-239.

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