

ISTANBUL TECHNICAL UNIVERSITY ★ GRADUATE SCHOOL

**PALEOCLIMATOLOGY AND GLACIAL GEOCHRONOLOGY OF THE
WESTERN TAURUS (SW TÜRKİYE) USING COMPARATIVE, RELATIVE
AND NUMERICAL METHODS**

Ph.D. THESIS

Onur ALTINAY

Department of Solid Earth Sciences

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İSTANBUL TEKNİK ÜNİVERSİTESİ ★ LİSANSÜSTÜ EĞİTİM ENSTİTÜSÜ

**KARŞILAŞTIRMALI, GÖRECELİ VE SAYISAL YÖNTEMLER IŞIĞINDA
BATI TOROSLAR'IN (GÜNEYBATI TÜRKİYE) PALEOKLİMATOLOJİSİ VE
BUZUL JEOKRONOLOJİSİ**

DOKTORA TEZİ

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To my mom,



FOREWORD

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ABBREVIATIONS

a.s.l.	: above sea level
AABR	: Area-Altitude Balance Ratio
AC	: Akdağ Cirque
ACP	: All control points
AMS	: Accelerator Mass Spectrometry
ANSTO	: Australian Nuclear Science and Technology Organization
AWOS	: Automatic weather observation station
C#	: Cirque number
CP	: Control point
DEM	: Digital elevation model
DJF	: December, January, February
DSM	: Digital surface model
EH	: Early Holocene
ELA	: Equilibrium line altitude
EPSG	: European Petroleum Survey Group
GaC	: Gelincikana Cirque
GC	: Gelincik Cirque
GIS	: Geographic information system
GM	: Ground moraine
GPS	: Global Positioning System
GRG	: Gedik rock glacier
HDPE	: High-density polyethylene
HGM	: Harita Genel Müdürlüğü
İC	: İkizler Cirque
ICE-D	: Informal Cosmogenic-nuclide Exposure-age Database
ICP-ES	: Inductively coupled plasma emission spectroscopy
ICP-MS	: Inductively coupled plasma mass spectrometry
IQR	: Interquartile range
KLV	: Kartal Lake Valley
KV	: Karadere Valley

L x W x H	: Length x Width x Height
LG	: Late-glacial
LGM	: Last Glacial Maximum
LH	: Late Holocene
LIA	: Little Ice Age
LiDAR	: Light detection and ranging
LKV	: Lower Kuruova Valley
LLM	: Left lateral moraine
LOI	: Loss of ignition
LSD/SF	: Lifton-Sato-Dunai / Sato Flux dependent
MAP	: Mean annual precipitation
MAT	: Mean annual temperature
MBCC	: Mountain-based calibration curve
MH	: Middle Holocene
MIS	: Marine Isotope Stage
NC	: Nivation cirque
NCAR	: National Center for Atmospheric Research
NCEP	: National Centers for Environmental Prediction
NGRIP	: North Greenland Ice Core Project
ORG	: Ortayazı rock glacier
OSL	: Optically stimulated luminescence
RLM	: Right lateral moraine
RM	: Recessional moraine
RRIM	: Red relief image map
SBCC	: Section-based calibration curve
SH	: Schmidt hammer
SHD	: Schmidt hammer (exposure-age) dating
SRD	: Schmidt hammer (relative-age) dating
SRTM	: Shuttle Radar Topography Mission
TC	: Toklu Cirque
TCND	: Terrestrial Cosmogenic Nuclide Dating
TL	: Thermoluminescence
TM	: Terminal moraine
TRI	: Terrain Ruggedness Index
TV	: Taşkuzluklu Valley

UAV	: Unmanned aerial vehicle
UKV	: Upper Kuruova Valley
UTM	: Universal Transverse Mercator
WGS84	: World Geodetic System 1984
X	: Test boulder
YD	: Younger Dryas





SYMBOLS

^{14}C	: Radioactive isotope of Carbon
^{36}Cl	: Radioactive isotope of Chlorine
$^{35}\text{Cl}, ^{37}\text{Cl}$	: Stable isotopes of Chlorine
^{10}Be	: Radioactive isotope of Beryllium
^3He	: Stable isotope of Helium
$^{21}\text{Ne}, ^{22}\text{Ne}$	: Stable isotopes of Neon
^{26}Al	: Radioactive isotope of Aluminium
$^{\circ}\text{C}$	: Degree Celsius
DD	: Decimal degrees
m/m^2	: meter/square meter
km/km^2	: kilometre/square kilometre
%	: Percent
~	: Tilde
$\pm$	: Plus and minus
ppm	: Parts per million
ka	: kilo annum
BP	: Before present
AD	: Anno domini
HNO_3	: Nitric acid
NaCl	: Sodium Chloride
AgNO_3	: Silver Nitrate
^{36}S	: Stable isotope of Sulfur
BaSO_4	: Barium sulfate
MV	: Megavolt
K_2O	: Potassium oxide
a	: Year
Ma	: Million year
Ca	: Calcium
g	: gram
K	: Potassium

^{40}Ca	: Stable isotope of Calcium
^{39}K	: Stable isotope of Potassium
Al_2O_3	: Aluminum oxide
CaO	: Calcium oxide
Fe_2O_3	: Iron oxide
MgO	: Magnesium oxide
MnO	: Manganese oxide
Na_2O	: Sodium oxide
P_2O_5	: Phosphorus pentoxide
SiO_2	: Silicon dioxide
TiO_2	: Titanium dioxide
CO_2	: Carbon dioxide
Sm	: Samarium
Gd	: Gadolinium
U	: Uranium
Th	: Thorium
Cl	: Chlorine
W/m^2	: watts per square meter
R^2	: coefficient of determination

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PALEOCLIMATOLOGY AND GLACIAL GEOCHRONOLOGY OF THE WESTERN TAURUS (SW TÜRKİYE) USING COMPARATIVE, RELATIVE AND NUMERICAL METHODS

SUMMARY

Türkiye possesses a remarkable heritage of the Late Pleistocene glaciations, evident across almost all of its lofty mountainous. Numerous individual mountains, particularly in the Eastern Black Sea and Taurus mountain ranges, host most of these glacial landscapes.

In the 20th century, many pioneering studies on Late Pleistocene glaciations were conducted by various researchers in Anatolia. However, it was not until the 21st century that a comprehensive understanding of the glaciers in both spatial and temporal dimensions became possible. Thanks to numerical dating techniques, we now know the temporal resolution of the glaciation. Among these methods, the Terrestrial Cosmogenic Nuclide Dating (TCND) has gained considerable popularity since the early 2000s. In this study, first I reviewed the existing records on glacial chronology in the SW Türkiye. Additionally, I employed TCND to elucidate and fill certain gaps in the glacial geochronology of the Western Taurus Mountains. Finally, I tested a calibrated dating technique known as Schmidt Hammer Exposure-Age Dating (SHD) on different lithologies in the region for a cross-validation of TCND ages.

In the first part of this thesis, the Late-glacial to Holocene glacier chronologies in Türkiye are reviewed. Despite the extensive extent of Turkish glaciers during the Last Glacial Maximum (LGM), traces of glaciers from the Late-glacial period are notably scarce in Türkiye. Certain landforms have been evaluated as Late-glacial and/or younger, but solid evidence, such as numerical dating, has not been obtained. In this section, Late-glacial and Holocene glacial areas are identified based on their TCND chronologies. Most of these areas are part of the Taurus Mountains, including the Dedegöl Mountains, Geyikdağ, Bolkar Mountains, and Aladağlar. Younger Dryas advances are evident in these mountains, especially in Aladağlar. Erciyes Volcano is the only location outside the Taurus Mountains in this study. Generally, the Late-glacial to Holocene glaciers of Turkish mountains extend from 14.8 ± 2.9 ka to 1.0 ± 0.3 ka, with certain Early to Late Holocene ages. Evidence for Little Ice Age (LIA) glacier advances is rare in Türkiye.

Despite the rich glacial chronology of the Taurus Mountains, there are still areas with well-developed glacial geomorphology but lacking a strong dating control. In this thesis, I examined two of these areas to reveal their glacial history. One of these areas is known as Mount Barla, situated in the northernmost part of the Western Taurus Mountains. Polygenetic and polycyclic processes play an important role in the origin of the glacial valleys of Mt. Barla. Glacial, karst, mass movement, and fluvial processes are intertwined, paving the way for glacial valley formations. The paleoglaciers here date back to 78.5 ± 17.6 ka (MIS5). It is also observed that there was a co-deglaciation between two different glacial valleys. The TCND ages indicate that these glacial retreats occurred around 42.1 ± 5.2 ka and 45.5 ± 6.8 ka (MIS 3).

Despite the absence of an LGM deglaciation, the final glacial retreat dates back to 16.6 ± 2.6 ka (Late-glacial).

Mount Davraz is another mountain with a well-established glacial geomorphology. However, this mountain has also lacked numerical ages until this thesis. It is located roughly 30 km south of Mount Barla, yet the paleoglaciers shaped this mountain's geomorphology in a notably different manner. The peculiar glacial geomorphology, considering its relatively small scale, is attributed to the existence of piedmont glaciers during the Late Pleistocene. The paleoglaciers within the cirques on the north-facing slopes, aligned in an E-W direction, were sufficient to support these piedmont glaciers. The paleoglaciers of Mount Davraz most probably benefited from strong southerly winds. These winds carried snow from the relatively flat top side to the north-facing slopes of the mountain. Consequently, the piedmont glaciers moved northward, extending over a northerly flat area and forming the Davraz paleoglacier. In this study, the remnants of this paleoglacier were dated using the TCND method. These remnants, known as "hummocky moraines," are common landforms formed after the disappearance of an ice sheet or a piedmont glacier. I utilized TCNs obtained from glacial boulders located on these moraines to calculate the latest deglaciation phase in Mount Davraz. The results indicate two periods: the Davraz paleoglacier began to retreat from east to west around 21.8 ± 2.4 ka (LGM) and completely disappeared around 17.7 ± 2.2 ka (Late-glacial).

Finally, I utilized Schmidt Hammer Exposure-Age Dating (SHD) to evaluate its applicability on Mount Sandıras, Karadağ, and Akdağ, as part of the Western Taurus Mountains. I measured rebound values (R-values) from the glacial surfaces to assess surface hardness. These measurements were conducted on both dated ("control surfaces" with TCND ages) and undated surfaces (test surfaces), with each control and test surface forming a sample group. The primary objective of this approach was to ascertain whether serpentinitized harzburgite (Mount Sandıras) and limestone (Karadağ and Akdağ) surfaces are suitable for SHD applications. While the method yielded partial success with Mount Sandıras samples, the results from Karadağ and Akdağ did not align with expectations. The SHD glacial chronology of Mount Sandıras exhibited a substantial fit with its TCND glacial chronology.

KARŞILAŞTIRMALI, GÖRECELİ VE SAYISAL YÖNTEMLER IŞIĞINDA BATI TOROSLAR'IN (GÜNEYBATI TÜRKİYE) PALEOKLİMATOLOJİSİ VE BUZUL JEOKRONOLOJİSİ

ÖZET

Yüksek ve dağlık bir ülke olmasına bağlı olarak, Geç Pleyistosen buzullaşmalarının izlerine Türkiye'nin dört bir yanında rastlamak mümkündür. Bu izlerin en geniş alana yayıldığı yerler ise Doğu Karadeniz ve Toros Dağları olarak kabul edilebilir. Bu dağ sıraları, ülkenin kuzeyinde ve güneyinde doğu-batı doğrultusunda uzanır ve içerisinde bir çok buzullaşmış zirve barındırır.

20. yy'nin başlarından itibaren Türkiye'de yer alan Geç Pleyistosen buzullaşma alanları üzerinde bir çok öncü çalışma ve bu çalışmaları takip eden detaylı çalışmalar yapılmıştır. Özellikle de 21. yy'nin başlarından itibaren, buzullaşma kronolojilerine dair yapılan araştırmaların sayısında ciddi bir artış görülmektedir. Bu araştırmalar sayesinde özellikle de paleoiklim çalışmaları için önemli bir veritabanı oluşturulmuştur. Söz konusu araştırmalarda kullanılan yöntemler arasında en popüler olanlardan biri ise kozmojenik izotoplarla yüzey tarihlendirmesi (TCND) yöntemidir. TCND çalışmaları, Türkiye'de kabaca 2000'li yılların başlarında başlamıştır. Geçtiğimiz çeyrek asır içerisinde Türkiye'de yapılan çalışmalar kapsamında da popüleritesi giderek artmıştır. Bu tez çalışmasında ise TCND yöntemi, tez kapsamındaki ilk üç çalışmanın ana metodolojisini oluştururken, son çalışmada ise Schmidt çekici yüzey tarihlendirmesi (SHD) yöntemine dair veri tabanı olarak kullanılmıştır. SHD yönteminde amaç, kayaların yüzey sertliklerini ölçmeye yarayan Schmidt çekici kullanılarak, elde edilen değerler (R-değeri), ölçüm yapılan kayaların TCND yaşları ile karşılaştırılır. Buna bağlı olarak oluşturulan kalibrasyon eğrisi sayesinde yaşa sahip olmayan kaya yüzeylerinin yüzeyleme yaşları, regresyon analizi ile hesaplanır. SHD yöntemi, bu tez kapsamında Türkiye'de ilk kez uygulanmıştır.

Türkiye, Son Buzul Maksimumu (SBM) buzullaşmalarını işaret eden zengin bir buzul kronolojisine sahip olsa da SBM sonrası (Geç-buzul ve Holosen gibi) için aynı şeyleri söylemek pek mümkün değildir. Tezin ilk kısmında, Türkiye'deki Geç-buzul'dan Holosen'e kadar olan buzul kronolojileri değerlendirilmiştir. Bu kapsamda, TCND kullanılarak tarihlendirilen buzullaşma yüzeyleri dikkate alınmıştır. TCND envanterine yer alan yaşlar, güncel kozmojenik izotop üretim hızları ve ölçekleme faktörleri kullanılarak tekrar hesaplanmıştır. Söz konusu kronolojiler incelendiğinde, Geç-buzul–Holosen periyodu içerisinde kalan bu alanların kabaca Toros Dağları üzerine gruplandığı görülmüştür. Ek olarak ise Orta Toroslar'ın hemen kuzeyinde, İç Anadolu'da yer alan Erciyes Volkanı'nın (ya da Erciyes Dağı) çalışma alanına dahil olduğu tespit edilmiştir. Toros Dağları'nda batıdan doğuya doğru Dedegöl Dağları, Geyikdağ, Bolkar Dağları ve Aladağlar çalışma alanındaki diğer dağlardır. Yapılan çalışmada, Geç-buzul'a dair hesaplanan yaşlar Geyikdağ (15.0 ± 3.0 bin yıl öncesi ve 14.8 ± 2.9 bin yıl öncesi) ve Aladağlar (14.2 ± 0.6 bin yıl öncesi) ile sınırlı kalmıştır. Dikkat çeken önemli noktalardan biri, Aladağlar'daki yedi örnek yaşından altısına tekabül eden Genç Dryas yaşlarıdır. Dedegöl Dağları ve Geyikdağ'da da Genç Dryas'a

ait iki yaş tespit edilmiştir. Bununla birlikte Erken Holosen yaşları da çalışma sahası içerisinde Aladağlar dışında yaygın bir şekilde görülen bir diğer periyodu temsil etmektedir. Orta Holosen yaşları yalnızca Geyikdağ'da görülürken, Erciyes Dağı, 1.0 ± 0.3 bin yıl öncesi olan Geç Holosen yaşı ile kronolojiyi sonlandırmaktadır. Bu tespitler elbette ki TCND kronolojileri baz alınarak yapıldığından, Türkiye'de farklı Geç-buzul-Holosen buzullaşma yüzeylerinin de bulunduğunu/bulunabileceğini belirtmek gerekir.

Toros Dağları, her ne kadar zengin bir buzul kronolojisine sahip olsa da hala tarihlendirmemiş olan buzullaşma alanları mevcuttur. Hatta, yeni buzullaşma alanlarının tespiti de güncel olarak devam etmektedir. Tezin ikinci ve üçüncü kısımları, sırasıyla Barla ve Davraz dağlarına dair yaptığım Geç Pleyistosen buzullaşmaları çalışmalarına ayrılmıştır. Her iki dağ da on yıllardır Türkiye buzul envanterinde yer almasına karşın buradaki buzullaşmalara dair bir kronoloji oluşturulmamıştır. Bu tez kapsamında TCND metodu kullanılarak söz konusu iki dağa ait ayrı buzul kronolojileri oluşturulmuştur. Ek olarak, Geç Pleyistosen buzullaşmaları jeomorfolojik açıdan da tartışılmış ve yorumlanmıştır.

Barla Dağı, Batı Toroslar'ın en içerde kalan kısmını oluşturur. Eğirdir Gölü'nün (916 m) batı kıyısından deniz seviyesinden 2798 m'ye kadar yükselir. Üzerinde kabaca kuzey yönlü olan Ortayazı, Gedik ve Kapıdere buzul vadileri yer alır. Bu vadiler, zirve çevresinde yer alan sirk buzullarının akarsu vadilerini aşındırması sonucu gelişmiştir. Tezin bu bölümünde, söz konusu vadiler polijenik ve polisiklik gelişim kapsamında ele alınmıştır. Ayrıca, Gedik ve Kapıdere vadilerindeki moren bloklarından alınan TCND örnekleri ile vadilerdeki buzullaşma evriminin geçmişi ortaya konmuştur. Ortayazı ve Gedik vadilerinde, Akdeniz çevresinde sık rastlanan buzul-karst (glasliyokarst) etkileşimi, bu vadilerin gelişimindeki temeli oluşturmaktadır. Zirve seviyesine yakın yer alan dolinlerin, sıklara dönüşmesine bağlı olarak söz konusu vadilerin geliştiği tahmin edilmektedir. Kapıdere Vadisi'nde ise durum biraz daha karmaşık bir hal almaktadır. Buradaki gelişimin ise önce bir heyelanla başladığı, daha sonra bu heyelanın taç kısmının içerisinde biriken karın buzula dönüşmesi ile devam ettiği ve sonuç olarak burada bir sirk gelişimiyle ilk aşamanın tamamlandığı düşünülmektedir. Sonrasında ise buradaki buzul ortadan kalktıktan sonra, paraglasiyal süreçlere bağlı olarak buradaki sirk duvarının göçtüğü ve buzul sonrası süreçlerin yoğun bir şekilde aktif olduğu ortaya konulmuştur.

Barla Dağı'nın, tıpkı jeomorfolojik evrimi gibi karmaşık olan bir buzul kronolojisine sahip olduğu düşünülmektedir. Buna göre, Kapıdere Vadisi'nin sağ yanal moreninden elde edilen TCND yaşı, 78.5 ± 17.6 bin yıl önce gerçekleşen bir buzul gerilemesini işaret etmektedir. Bu da denizel izotop seviyelerinin beşincisine denk gelmektedir (*ing. Marine Isotope Stage; MIS 5*). Farklı kanıtlarla doğrulanması şartı ile birlikte bu yüzeylenme yaşı, Türkiye'de şu ana kadar tespit edilen en yaşlı buzul gerilemesini işaret etmektedir. Yine, Kapıdere Vadisi ve Gedik Vadisi'nden elde edilen cephe moreni yaşları, 42.1 ± 5.2 ve 45.5 ± 6.8 bin yıl öncesi ile *MIS 3*'ü temsil etmektedir. Son olarak ise, Gedik Vadisi'nin sol yanal moreninin yaşı (16.6 ± 2.6 bin yıl önce), Barla Dağı'ndaki son buzul gerilemesini Geç-buzul olarak işaretlemiştir. Burada SBM'ye dair bir gerileme tespiti yapılamamakla birlikte, Geç-buzul morenine göre daha geride ve yüksekte yer alan ve yaşı olmayan bir morenin de morfostratigrafiye bağlı olarak daha genç bir gerilemeyi işaret ettiği düşünülmektedir.

Davraz Dağı ise Barla Dağı'nın 30 km güneyinde yer alan fakat tamamıyla farklı bir jeomorfolojiye sahip bir başka Geç Pleyestosen buzullaşma alanıdır. Dağın en yüksek zirvesi deniz seviyesinden 2637 m yükseklikte yer alır. Barla Dağı'nın aksine geniş bir plato tipi zirve düzlüğüne sahiptir. Buradaki buzul jeomorfolojisinin gelişimi dağın kuzeye bakan yamaçlarında, doğu-batı yönlü sıralanmış olan 10 adet sirk ile başlamıştır. Topografya ve yapısal unsurlar nedeniyle iyi gelişmiş buzul vadilerine rastlanmamaktadır. Geç Pleyistosen'de sirkler içerisinde gelişmiş olan buzullar, yamaçlardan aşağıya doğru hareket etmiş, yine plato yapısında düz bir alana ulaştıktan sonra burada tıpkı bir alüvyal yelpaze gibi yayılmışlardır. Bu şekilde oluşan buzullar dağ eteği yani "piedmont" buzulları olarak adlandırılırlar. Bu piedmont buzulları sayesinde, plato üzerinde dağın kütleline nazaran çok geniş bir çeşit buzul örtüsü oluşmuştur. Bu buzul örtüsü ortadan kalktıktan sonra tümseksi moren olarak adlandırılan çok büyük oranda kaotik olan yeryüzü şekillerden oluşan geniş bir arazi ortaya çıkmıştır. Barla Dağı'na ve çevresindeki diğer dağlara göre görece küçük ölçekli olan Davraz Dağı'nın, Geç Pleyistosen buzullaşmalarından yoğun bir şekilde etkilendiği söylenebilir. Bu noktada dönemin sıcaklık ve yağış koşullarına ek olarak, güney yönlü esen güçlü rüzgarların, dağın zirve kısımlarından taşıdığı karların da buzul gelişimini pozitif etkilediği düşünülmektedir.

Davraz Dağı'nın buzul kronolojisi ise Toroslar ve Türkiye genelindeki SBM ile uyum içerisinde görünmektedir. Tümseksi morenlerin üzerinde yer alan altı moren bloğunun tarihlendirmesine bağlı olarak buradaki son buzul gerilemesinin 21.8 ± 2.4 bin yıl önce (SBM) başlayarak, kabaca 17.7 ± 2.2 bin yıl önce (Geç-buzul) sonlandığı ve buzulun ortadan kalktığı ortaya konmuştur. Buna göre Davraz paleobuzulu, tümseksi moren alanının daha alçak olduğu doğu kısmından başladığı tespit edilmiştir. Buradaki morfoloji ayrıca, paleobuzulun erimeye başladığı noktada buzulaltı erime suyunun drenaja açıldığı yeri işaret etmektedir. Batıdan başlayıp bölgesel bir erime deseni takip eden Davraz paleobuzulun gerisinde bıraktığı malzeme, doğuya doğru gittikçe artmaktadır. Doğuda daha kalın olan tümseksi morenler bunu göstermektedir. Paleobuzulun erimesi ilk başlarda ölü buzul tipinde gerçekleşirken, yamaca yakın alanlarda ise Davraz paleobuzulu ve sirk buzullarının bağlantısının kopmasıyla, yamaçlarda gerileme şeklinde devam etmiştir. Tahminen geç Geç-buzul periyodunda tüm buzullar ortadan kalkmıştır.

Tezin son kısmında ise Schmidt çekici yüzey tarihlendirmesi (SHD) yöntemi, harzburjit ve kireçtaşı litolojileri üzerinde test edilmiştir. Çalışma alanı olarak Batı Toroslar'ın en batısında yer alan Sandıras Dağı (harzburjit litoloji), Karadağ ve Akdağ (kireçtaşı litolojileri) seçilmiştir. Çalışma sahalarında TCND yaşlarına sahip bloklar "kontrol yüzeyi" olarak, herhangi bir yaşa sahip olmayan bloklar ise "test yüzeyi" olarak adlandırılmıştır. Her bir test yüzeyi, kontrol yüzeyine mümkün olduğunca yakın seçilmeye çalışılmıştır. Bu stratejide, SHD yönteminin harzburjit ve kireçtaşı litolojileri üzerindeki başarısını test etmek temel hedef olarak belirlenmiştir. Sandıras Dağı'ndaki sonuçlar kusursuz olmasa da tatmin edici olmuştur. 0.66 ve 0.92 gibi korelasyon katsayılarına ulaşılan kalibrasyon eğrilerine bağlı olarak göree tutarlı SHD kronolojileri elde edilmiştir. Diğer yandan, Karadağ ve Akdağ'daki TCND yaşları ve R-değerleri arasında tutarlı bir ilişki tespit edilememiş ve buna bağlı olarak kalibrasyon eğrileri oluşturulamadığı için SHD yaşları hesaplanamamıştır.



1. INTRODUCTION

Glaciers, responsive indicators of climate fluctuations, continually expand and contract in reaction to variations in temperature, snowfall, and other environmental factors (Benn & Evans, 2010). In essence, the development of glaciers requires optimal environmental conditions. On a global scale, the decisive element in meeting these conditions is the high latitude factor (i.e., polar regions), while locally, glacier development is primarily controlled by high-elevation areas (i.e., mountains). In the absence of these essential conditions, it cannot be assumed that glacier development occurs; instead, glaciers may retreat or disappear completely. The last and the most effective period in which the necessary conditions for the formation of glaciers were provided on a global scale corresponds to the end of the Late Pleistocene, known as the Last Glacial Maximum (LGM) (from 26.5 to 19 ka ago; Clark et al., 2009).

Glaciers can serve as proxies in climate studies due to their high-order correlation with the climate. For instance, the composition and macrostructure of ice extracted from extensive ice sheets such as Antarctica and Greenland, as well as from mountain glaciers, offer insights into regional air temperatures and the global/regional atmospheric circulation and composition (Goudie, 2003). Moreover, paleoclimatic inferences can be drawn from glaciers that no longer exist. By combining the results obtained from the evidence provided by paleoglacial formations, it is possible to make inferences about paleoclimate conditions. This helps interpret not only present and future natural climate variability but also the anthropogenic impacts on the current climate state.

In the Eastern Mediterranean, where the presence of active glaciers is marginal today, there exists a rich glacier chronology extending from the pre-LGM to the Holocene. Hughes & Woodward (2017) stated that recent dating studies on Mediterranean glaciations have particularly focused on Iberia/The Pyrenees and Türkiye. Starting from the second half of the 2000s, with an increasing number of terrestrial cosmogenic nuclide dating (TCND), a comprehensive glacier chronology has been established for Quaternary glaciations in the Eastern Mediterranean. In Türkiye, these studies have

roughly focused on three main regions: (1) the Northeastern Black Sea part, which is known as the Pontic Mountains in the Eastern Black Sea region, and (2) the Southern and Southwestern Taurus region, which lies along the Mediterranean coastal region, and lastly (3) individual volcanoes/mountains spread over Anatolia. The studies conducted in the Eastern Black Sea region can be listed chronologically as follows: Mt. Kaçkar (Akçar et al., 2007); Mt. Verçenik (Akçar et al., 2008), Başyayla Valley (in Eastern Black Sea Mountains) (Reber et al., 2014, 2022), Mt. Karçal (Dede et al., 2017). Almost all of the studies conducted in the Eastern Black Sea region (except for Mt. Karçal by Dede et al., 2017) are located on the Eastern Black Sea Mountains. On the other hand, the studies conducted in the Mediterranean Taurus Mountains are as follows: Mt. Sandıras (Sarıkaya et al., 2008), Mt. Dedegöl (Zahno et al., 2009; Köse et al., 2019), Aladağlar (Zreda et al., 2011), Akdağ (Sarıkaya et al., 2014), Bolkar Mountains (Çiner & Sarıkaya, 2017), Geyikdağ (Çiner et al., 2015, 2017; Sarıkaya et al., 2017), Mt. Barla (Altınay et al., 2022), Mt. Karanfil (Köse et al., 2022), Mt. Davraz (Altınay et al., 2024). Other studies on individual mountains, such as Mt. Erciyes (Sarıkaya et al., 2009) and Uludağ (Zahno et al., 2010; Akçar et al., 2014), are located in the northwest and central regions of Türkiye, respectively. On Mt. Lebanon (Moulin et al., 2022), in the easternmost part of the Mediterranean, there are also paleoglaciatic studies by TCND methods exist.

It is known that the ages obtained within the scope of these studies cluster in the last glaciation, especially in between the boundaries of the pre-LGM, the Last Glacial Maximum (LGM), and the Late-glacial periods. In addition to TCND studies, ages of glacial landforms obtained with optically stimulated luminescence (OSL) are also available in the Eastern Mediterranean (e.g., Çilgın, 2015; Bayrakdar et al., 2017). A significant portion of the dating studies on Quaternary glaciations in the Eastern Mediterranean has been conducted on the Western and Central Taurus Mountains (hereafter referred to as the Mediterranean Taurus), extending across the southern coasts of Türkiye. Despite the presence of active glaciers and well-developed glacial landforms in the Southeastern Taurus Mountains, which are part of the broader Taurus Mountain range, no dating studies have been conducted in this area. The only exception is the study by Yeşilyurt (2017) in the Kavuşşahap Mountains (3634 m a.s.l.).

The Taurus Mountains, forming part of the Alpine-Himalayan orogenic system, and its Mediterranean part extends between the sea (Mediterranean) and the interior regions of Anatolia, and acts as a natural barrier. While they prevent moisture from reaching the interior parts of Anatolia, they cause the majority of precipitation to concentrate in the coastal region, resulting in a significant reduction in rainfall when clouds pass over the mountains and reach Central Anatolia (Şensoy, 2008). On the other hand, the southeastern part of the Taurus Mountains, although influenced by the continental climate conditions, also indirectly experiences the influence of the Mediterranean climate. As an illustration of this phenomenon, the area is impacted (e.g., with increased precipitation) by the intensity and positioning of the "Cyprus cyclones" originating from the Mediterranean Sea (Litt et al., 2014). While the number of Quaternary glacier dating studies in the Mediterranean Taurus is relatively high, it cannot be said that these studies have been conducted in every region where glacial traces are observed. Indeed, new glaciated areas in the Taurus Mountains have continued to be discovered even in recent years (e.g., Bayrakdar et al., 2020, 2023).

Glaciers currently present in the Taurus Mountains are located in the high-altitude regions of the range, similar to the active glaciers found elsewhere in Türkiye (cf. Sarıkaya & Tekeli, 2014; Azzoni et al., 2020; Keserci et al., 2023). Presently, apart from relatively small active cirque glaciers on Geyikdağ (Keserci et al., 2023), the existence of active glaciers cannot be mentioned in the Western Taurus Mountains. However, in the Central Taurus Mountains, the presence of debris-covered glaciers can be noted in Aladağlar and Bolkar Mountains (Gürgen et al., 2010; Çalışkan et al., 2012; Çiner & Sarıkaya, 2017; Köse et al., 2021). Contrary to the Mediterranean Taurus, active glaciers are present in the Southeastern Taurus Mountains, including Buzul (Cilo) and İkiyaka mountains (Erinç, 1952; Erinç, 1953; Kurter, 1991; Yeşilyurt et al., 2013; Sarıkaya & Tekeli, 2014). In this region, Mt. Buzul (Cilo) also possesses the second-largest glacier area (3.82 km²) in Türkiye, after Mt. Ağrı (Ararat) (Sarıkaya & Tekeli, 2014; Azzoni et al., 2017, 2020; Azzoni et al., 2022).

Current TCND studies indicate that Late-glacial and Holocene glaciations after the Last Glacial Maximum (LGM) are confined to the Mediterranean Taurus and Mt. Erciyes in Central Anatolia. In **chapter 2**, five mountains in Türkiye are reviewed, where Late-glacial to Holocene chronologies are present. These glacial chronologies are situated in the western and central Taurus Mountain Ranges, as well as in central

Anatolia. The timing of the glaciations has predominantly been determined through cosmogenic ^{36}Cl surface exposure dating. Some of the previously published cosmogenic ages underwent recalculation using updated production rates and scaling schemas. Late-glacial and Younger Dryas ages were reported.

Although the Taurus Mountains display numerous glacier chronologies, significant glaciated areas within the region remain undated. Recently, glacial landforms in the Barla and Davraz mountains, previously undated in the Western Taurus section, were dated using the TCND method in this thesis study and added to the Eastern Mediterranean glacier inventory. Both of these mountains are situated in a transitional climate zone between Mediterranean and continental Anatolian climate. Türkeş et al. (1996) classified this zone as a rainfall region and termed it as the Mediterranean transition (MEDT). On the other hand, Atalay & Efe (2010), delineated the boundaries of the same section based on forest ecosystems and referred to it as the Mediterranean transitional region. Barla and Davraz mountains are located within both of the defined areas. This transitional zone is characterized by moderate rainfall in winter and spring, with lighter precipitation in summer (Türkeş et al., 1996; Duzenli et al., 2018). The precipitation regime in this zone is categorized as transitional due to its dependence on the combined effects of frontal and convective rainfall (Sarış et al., 2010). As the rain shadow effect reduces the mean annual precipitation to 400 mm, the average yearly temperature in the lowlands ranges from approximately 10 to 13 °C, with the average winter temperature staying above freezing (Atalay & Efe, 2010).

In **chapter 3**, Quaternary glaciations on Mt. Barla were examined from a geomorphological and geochronological perspective. The periods of glaciation and their geomorphological impacts were discussed based on the retreat ages of glaciers obtained from glacial landforms by TCND. Additionally, the geomorphological evolution between the pre-glacial and post-glacial periods was discussed within the framework of polygenic and polycyclic processes. The evolution of glacial valleys in Mt. Barla was individually studied, and detailed geomorphological maps were produced. The geomorphological development related to the relationship between karstification-glaciation and mass movements-glaciation, influenced by polygenic evolution, was also discussed.

Chapter 4 consists of the glacial- geomorphology and chronology of Mt. Davraz. The study area is located roughly 30 km south of Mt. Barla. It has a completely different

glacial geomorphology restricted by the structural controls in the area. This area represents a rare glacial environment for a mid-latitude region, with its vast expanse of hummocky moraines -especially concerning its size- and aligned cirques along with its north-facing slopes. The possibility that the development of these hummocky moraines is related to piedmont glaciers adds further significance to the Late Pleistocene glaciations in this area. In this section, a comprehensive geomorphological map of Mt. Davraz has been produced. Using the TCND method, the retreat ages of glaciers have been calculated based on boulders from hummocky moraines. Additionally, emphasis has been placed on the influence of topographic and local factors affecting glaciation in this area, and the relationship between glacial chronologies in nearby mountains such as Barla and Dedegöl has been examined. Finally, comparisons have been made with other piedmont glaciations and hummocky moraine areas in the Mediterranean region.

Chapter 5 presents preliminary results derived from a relative dating study based on TCND glacier chronologies in the Western Taurus Mountains. This study was conducted using the Schmidt hammer exposure-age (SHD) method, which has gained popularity in the past quarter-century. Representing the first SHD study on glacial landforms conducted in Türkiye, this section utilizes results from Schmidt hammer measurements taken at Mt. Sandıras, Akdağ, and Karadağ to relatively date previously undated moraine blocks in these areas. Calibration curves for relative dating were generated based on the use of TCND ages as reference points in the study. Additionally, attention was given to the challenges encountered when using the SHD method on carbonate rocks. In summary, it was tested whether the SHD method, which offers a much faster and cheaper alternative compared to other dating methods, would be effective in the Taurus Mountains.



2. LATE-GLACIAL TO HOLOCENE GLACIERS IN THE TURKISH MOUNTAINS¹

2.1 Introduction

With several mountain ranges >3000 m above sea level (a.s.l.), Türkiye is considered a mountainous country. The Taurus Mountains along the Mediterranean coast in the south and the eastern Black Sea Mountains along the northern coast of Türkiye constitute the most important mountain ranges in Türkiye. Apart from these mountain ranges, individual volcanoes such as Mt. Ağrı (Ararat) (5137 m a.s.l.) and Mt. Erciyes (3917 m a.s.l.) are also lofty mountains where glaciers and glacial deposits are encountered (Çiner, 2004).

Türkiye has remarkable glacial geomorphology inherited from the Late Quaternary glaciations. The presence of previous glacial areas was first mentioned in various European voyagers' travel notes (e.g., Ainsworth, 1842; Hamilton, 1843; Koch, 1846; Palgrave, 1872). However, the first scientific studies were undertaken by Maunsell (1901) in the Buzul Mountains of southeastern Türkiye and by Penther (1905) on Erciyes Volcano in central Anatolia. Later, most of these glacial and paleoglacial areas have been studied by various researchers between 1930 and 1950s (e.g., Krenek, 1932; Bobek, 1940a, b; Erinç, 1952) and then from the 1980s to present (e.g., Kurter & Sungur, 1980; Kurter, 1991; Çiner, 2004; Akçar et al., 2007, 2008; Sarıkaya et al., 2008, 2017; Bayrakdar, 2012; Cılığın, 2012, 2015; Sarıkaya, 2012; Çiner et al., 2015, 2017; Bayrakdar et al., 2017a; Dede et al., 2017; Köse et al., 2019). There are also several review studies about the glaciations in Türkiye (e.g., Erinç, 1952; Birman, 1968; Çiner, 2004; Sarıkaya et al., 2011, 2017; Sarıkaya & Çiner, 2015, 2017). Today, the largest Turkish glaciers are situated in the eastern part of the country (Azzoni et al., 2020). They are usually considered remnants of more extensive older glaciations (Erinç, 1952), from which they retreated to high-elevation cirques and slopes where

¹ This chapter is based on the paper "Altınay, O., Sarıkaya, M. A., & Çiner, A. (2020). Late-glacial to Holocene glaciers in the Turkish mountains. *Mediterranean Geoscience Reviews*, 2, 119-133."

climate, orographic factors, and slope orientation combine favorably to sustain them (Sarıkaya et al., 2014).

Thanks to the cosmogenic surface exposure dating of glacial landforms, Late Pleistocene glacial deposits in Türkiye now constitute one of the best-dated records of its kind in the world (Hughes & Woodward, 2017). These ages are scattered from pre-Last Glacial Maximum (pre-LGM) to Late Holocene with the majority of ages indicating ages that coincide with global LGM centered to c. 21 ka. Here, we focus specifically on reviewing Holocene glaciations that were dated by terrestrial cosmogenic nuclides, mainly by ^{36}Cl . Even though the eastern Taurus Mountain Range contains the highest peaks in the country, the absence of cosmogenic ages precludes our knowledge of any Holocene glacial activity. Therefore, available Holocene ages are only observed in the western and central Taurus Range. We recalculated the ages obtained from the earlier studies by the most recent production rates and scaling factors and presented a holistic view of the Holocene glaciation in the Turkish mountains. We assessed not only the Holocene ages but also selected ages ranging from Late-glacial to Younger Dryas, as this would help draw a useful conclusion on the onset of Holocene glaciations and provide comparisons between regions.

2.2 Study Area

The Taurus Mountain Range stretches 800 km from west to east and 150 km south to north along the Turkish coast of the Mediterranean Sea with altitudes reaching close to c. 3500 m a.s.l. The Taurus Mountains can be divided into three sectors; the western, central, and eastern Taurus, where several mountains were glaciated in the past. All reported Holocene glacial deposits were found on the western and central Taurus Mountain Range and one on the highest volcano in central Anatolia (Figure 2.1). From west to east, these are Dedegöl Mountains in the western Taurus Range, Geyikdağ, Bolkar, and Aladağlar in the central Taurus Range, and Mt. Erciyes (i.e., Erciyes Volcano) in central Türkiye (Table 2.1). These mountains are usually >2500 m a.s.l. in altitude, and their peak elevations are rising from west to east. The highest one is Mt. Erciyes, which is a stratovolcano, reaching 3917 m a.s.l. in elevation, and located in the Cappadocia Volcanic Province.

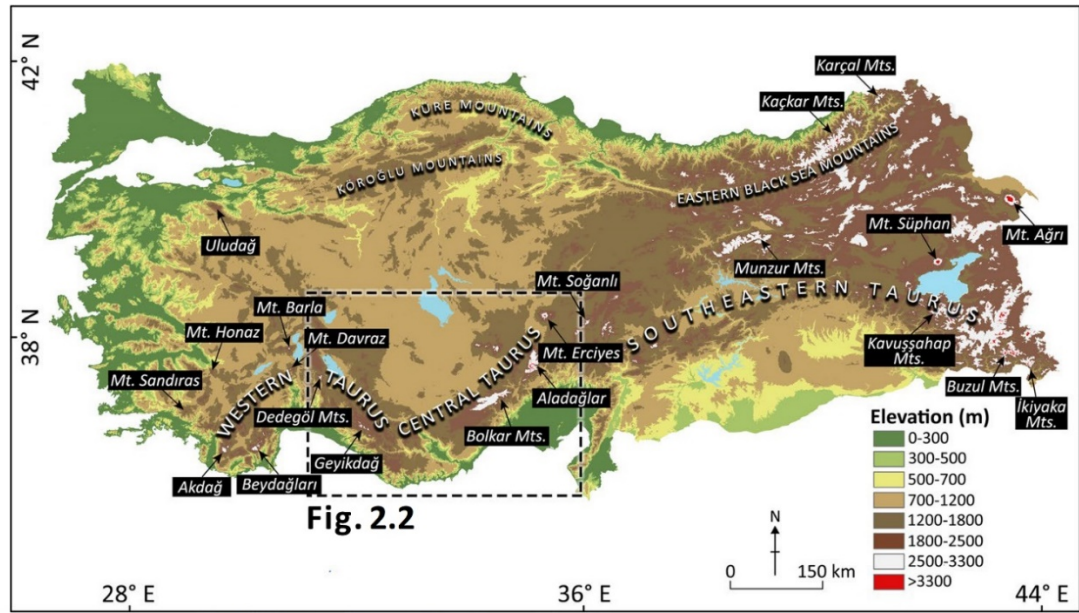


Figure 2.1 : Map of Türkiye showing the major glaciated mountains during the Late Quaternary.

Table 2.1 : List of the glaciated mountains that contain Late-glacial to Holocene moraines.

#	Mountain Name	Valley names	Peak Elevation	Latitude (DD°N)	Longitude (DD°E)	LGM ELA (m)	Modern ELA	Reference
1	Dedegöl Mts.	Karagöl & Kisbe	2992	37.6437	31.2835	2350-2400	3300-3500	Köse et al., 2019
2	Geyikdağ	Namaras & Susam	2850	36.8075	32.2021	2500	3400	Çiner et al., 2015
3	Geyikdağ	S. Çimi & N. Çimi & Güneycik	2850	36.8075	32.2021	2500	3400	Sarıkaya et al. 2017
4	Geyikdağ	Southern Güneycik	2850	36.8075	32.2021	2500	3400	Çiner et al., 2017
5	Bolkar Mts.	Karagöl & Alagöl & Elmalı	3524	37.3862	34.6087	2650	3450-3700	Çiner and Sarıkaya, 2017
6	Aladağlar	Hacer	3756	37.8366	35.1453	2700	3450	Zreda et al., 2011
7	Mt. Erciyes	Aksu & Üçker	3917	38.5318	35.4469	2700	3550	Sarıkaya et al., 2009

2.2.1 Geology

The Taurus Mountains constitute a substantial part of the southern and eastern Alpine orogenic belt (Özgül, 1976). This range is a diverse aggregation of rocks from the Late Precambrian crystalline basement, a mixed clastic-carbonate Paleozoic succession, and an Upper Triassic to Upper Cretaceous carbonate series (Okay, 2008). In the mountains where Late Quaternary glaciers developed, landscapes made up of limestones are widespread, and hence excellent examples of doline development and glaciokarst are observed (Çılğın et al., 2014; Öztürk et al., 2018).

To the north of the central Taurus Range and within the Cappadocia Volcanic Province, Mt. Erciyes (3917 m a.s.l.) was formed in two major stages of evolution (Şen et al., 2003; Kuzucuoğlu, 2019). The first stage began with basaltic lava flows and ended with substantial ignimbritic eruptions c. 3 Ma ago (Innocenti et al., 1975). The

second stage included basaltic, andesitic, dacitic, and rhyolitic lavas and ended pyroclastic eruptions and avalanches of debris in the Early Holocene (Sarıkaya et al., 2019).

2.2.2 Climate

The Anatolian climate is considered a transition between North Atlantic polar and African-Mediterranean tropical cyclonic air masses (Bozkurt & Şen, 2011; Macklin et al., 2002). The Anatolian Peninsula lies at the intersection of three major atmospheric systems: (1) main middle to high-latitude westerlies to the north and northwest, (2) mid-latitude subtropical high-pressure systems that generally extend from the Atlantic across the Sahara, and (3) monsoon climates of the Indian subcontinent and East Africa (Lionello et al., 2006; Akçar et al., 2007). During the winter, the region is affected by the strong thermal Siberian high-pressure systems (Wigley & Farmer, 1982). Because of the high-altitude plateau and the presence of mountains running parallel to the sea, the interior regions have a pronounced continental climate (Ünal et al., 2003). The coastal part of the Taurus Range is characterized by a Mediterranean climate with dry/hot summers and wet/temperate winters (Türkeş & Erhat, 2008). The air dries substantially off on the leeward side of the Taurus Mountains and becomes drier on the interior of Anatolia.

The Late Pleistocene glaciations show a broad distribution in Türkiye. On the contrary, the Holocene is a warmer epoch with important shifts in the temperatures. Nevertheless, some of the glaciers survived throughout the Holocene in Türkiye and nearby areas (e.g., Sarıkaya & Çiner, 2017; Leontaritis et al., 2020). It has been estimated that the air temperatures were between 2.1 °C and 4.9 °C lower than today and the precipitation ratio was up to two times more on Mt. Erciyes at the beginning of the Holocene (Sarıkaya et al., 2009).

Variation of today's equilibrium line altitudes (ELAs) is between 3400 m a.s.l. (on the coastal regions) and 4000 m a.s.l. (in the interior) (Sarıkaya & Çiner, 2017). Only a few mountains in eastern Türkiye are higher than the modern ELA (Azzoni et al., 2020). There are a few small glaciers (some debris covered) on Mt. Erciyes (Sarıkaya & Tekeli, 2014), on Bolkar Mountains (Çalışkan et al., 2014) and Aladağlar (Gürgen et al., 2010). The LGM ELA distribution was not different from the present, but 900-1500 m lower in altitude. The coastal mountains had paleo-ELAs at around 2000-2400

m a.s.l., while they increased to 2700 m a.s.l. towards the interior (Sarıkaya & Çiner, 2017).

2.3 Methodology

Since the publication of pioneering studies (Klein et al., 1982; Craig & Poreda, 1986; Kurz, 1986; Nishiizumi et al., 1986; Phillips et al., 1986; Marti & Craig, 1987) about cosmogenic ^3He , ^{21}Ne , ^{22}Ne , ^{10}Be , ^{26}Al and ^{36}Cl in terrestrial rocks, the methodological developments and their applications to earth surface sciences have become widespread (Dunai, 2010). The cosmogenic surface exposure dating method has been used for over thirty years in earth sciences, but its usage in Türkiye started towards the end of the 2000s. Since then, the number of studies with cosmogenic isotopes rapidly increased (e.g., Akçar et al., 2007, 2008, 2014; Sarıkaya et al., 2008, 2009; Sarıkaya, 2009; Zahno et al., 2010; Zreda et al., 2011; Reber et al., 2014; Çiner et al., 2015; Çiner & Sarıkaya, 2017; Köse et al., 2019).

All glacial chronologies in this review are based on cosmogenic ^{36}Cl that were compiled from the published dates from the literature. Later, we recalculated the ages by up-to-date production ^{36}Cl rates (Borchers et al., 2016; Marrero et al., 2016a; Phillips et al., 2016). We used the CRONUS Earth Web Calculator version 2.0 (Marrero et al., 2016b). Twenty-eight moraine ages (derived from 97 individual samples) from 14 valleys in 5 mountains were used for establishing the Late-glacial to Holocene glacial chronology of Türkiye (Tables 2.2 and 2.3).

Table 2.2 : The glaciated mountains have numerical age control.

#	Mountain / Valley name	Nuclide	Number of samples	Used samples	Outliers	Used samples by Interpreted ages	All samples by type including outliers
1	Dedegöl Mts. / Karagöl & Kisbe	^{36}Cl	9	5	4	EH=5	9 moraine boulders
2	Geyikdağ/ Namaras & Susam	^{36}Cl	12	11	1	LG to EH=4, LG to MH=5, YD=3	12 moraine boulders
3	Geyikdağ / S. Çimi & N. Çimi & Güneycik	^{36}Cl	15	14	1	EH=3, MH=11	15 moraine boulders
4	Geyikdağ / Southern Güneycik	^{36}Cl	4	4	0	MH=4	3 moraine boulders, 1 amalgamated boulder chips 3 outwash samples,
5	Bolkar Mts. / Karagöl & Alagöl & Elmalı	^{36}Cl	13	10	3	EH=7, MH=3	4 erratics, 6 moraine boulders
6	Aladağlar / Hacer	^{36}Cl	22	22	0	LG=3, YD=19	2 bedrock samples, 20 moraine boulders
7	Mt. Erciyes / Aksu & Üçker	^{36}Cl	21	20	1	EH=12, LH=8	5 outwash samples, 16 moraine boulders

Table 2.3 : Summary of TCND ages for the studied mountains.

#	Mountain name	Location	Moraine	Age (ka)	Reference
1	Dedegöl Mts.	Kisbe Valley	YD	11.5 ± 0.8	Köse et al., 2019
		Karagöl Valley	EH	9.8 ± 1.4	
2	Geyikdağ	Susam Valley	YD	12.7 ± 1.6	Çiner et al., 2015
		Namaras Valley	LG	15.0 ± 3.0	
			LG	14.8 ± 2.9	
		Güneycik Valley	EH	9.5 ± 0.5	Sarıkaya et al., 2017
			MH	5.9 ± 0.2	
		South Çimi Valley	MH	7.3 ± 0.3	
		North Çimi Valley	MH	8.0 ± 0.6	Çiner et al., 2017
		Southern Güneycik Valley	MH	5.8 ± 0.3	
3	Bolkar Mts.	Karagöl Valley	EH	9.0 ± 0.8	Çiner & Sarıkaya, 2017
			EH	8.7 ± 0.8	
		Alagöl Valley	EH	9.0 ± 0.4	
		Elmalı Valley	YD	12.7 ± 1.4	
			EH	9.3 ± 1.0	
4	Aladağlar	Hacer Valley	YD	12.3 ± 0.6	Zreda et al., 2011
			YD	12.5 ± 0.6	
			YD	12.3 ± 0.6	
			YD	13.1 ± 0.8	
			YD	13.1 ± 0.5	
			YD	13.5 ± 0.7	
			LG	14.2 ± 0.6	
5	Mt. Erciyes	Aksu Valley	LH	1.0 ± 0.3	Sarıkaya et al., 2009
			LH	2.8 ± 0.5	
			LH	4.1 ± 0.7	
			EH	8.3 ± 1.6	
			EH	9.5 ± 1.8	
		Üçker Valley	EH	10.2 ± 0.9	

2.4 Extent and Timing and Holocene Glaciers in Türkiye

Holocene paleoglaciers are distributed in the western and central Taurus Mountains, except in Mt. Erciyes, which is a volcano located in central Türkiye (Figure 2.2). Readers can refer to Sarıkaya et al. (2011), Akçar et al. (2015), and Sarıkaya & Çiner (2015, 2017) for details on other regions where Late Pleistocene glaciations were observed in Türkiye.

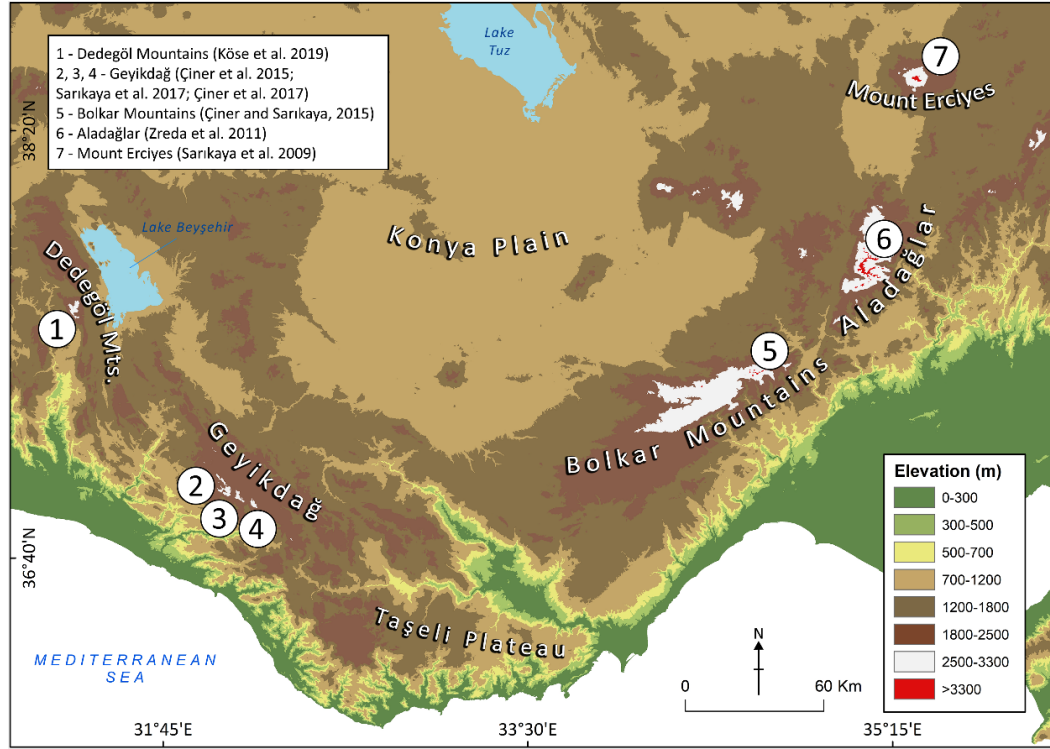


Figure 2.2 : Digital Elevation Model (DEM) of south-central Türkiye showing the glaciated mountains during the Late-glacial and Holocene.

There were many glaciated mountains in the Taurus Mountain Range (e.g., Mt. Sandıras (2295 m a.s.l.; Doğu, 1993; Sarıkaya et al., 2008), Mt. Honaz (2571 m a.s.l.; Yalçınlar, 1954; Erinç, 1955), Mt. Akdağ (3016 m a.s.l.; Doğu et al., 1999; Bayrakdar et al., 2017b), Mt. Karadağ (2418 m a.s.l.; Bayrakdar et al., 2017a; Altınay et al., 2019), Beydağları Mountains (3086 m a.s.l.; Louis, 1944; Messerli, 1967), Mt. Barla (2800 m a.s.l.; Ardos, 1977a; Monod, 1977), Mt. Davraz (2637 m a.s.l.; Ardos, 1977b; Monod, 1977), Mt. Soğanlı (3075 m a.s.l.; Yalçınlar, 1951; Ege & Tonbul, 2005), Kavuşşahap Mountains (3503 m a.s.l.; Sarıkaya & Tekeli, 2014; Yeşilyurt et al., 2018), Buzul Mountains (4135 m a.s.l.; Bobek, 1940a, b; Kesici, 2005) and İkiyaka Mountains (3794 m a.s.l.; Erinç, 1952). Nevertheless, only some of them have glacial

chronologies that give information about Late Quaternary glaciations. Here, we describe all Holocene and to some extent Late-glacial and Younger Dryas glacial deposits that have been dated by ^{36}Cl cosmogenic surface exposure method in the Taurus Range from the west towards the east and from Erciyes Volcano (Figure 2.2, Table 2.3).

2.4.1 Dedegöl Mountains

Dedegöl Mountains ($37^{\circ} 39' 22''$ N, $31^{\circ} 17' 36''$ E, 2992 m a.s.l.) (Figure 2.2) are located near the provincial borders of the cities of Isparta and Konya and are aligned in a NW-SE direction. The region's most dominant geological units are Mesozoic limestones, which make up the western Taurus Mountains' highest massif (Dumont & Monod, 1976).

Dedegöl Mountains constitute the westernmost region where Holocene glaciations are observed in Türkiye and have six glacial valleys (Sayacak, Kisbe, Karagöl, Karçukuru, Elmadere and Muslu valleys). Zahno et al. (2009) presented ^{10}Be and ^{26}Al surface exposure ages in Muslu Valley, and Köse et al. (2019) measured ^{36}Cl ages on the N-NE-trending Sayacak, Kisbe, and Karagöl valleys. Zahno et al. (2009) reported glaciers advanced no later than 24.3 ± 1.8 ka ago and provided evidence of major glaciation before the global LGM (21.0 ± 2.0 ka). They also indicated that ice-free conditions in the Muslu Valley occurred no later than 13.3 ± 1.5 ka ago. On the other hand, Köse et al. (2019) have reported an Early Holocene moraine dated to 9.8 ± 1.4 ka in the Karagöl Valley (Figure 2.3). Köse et al. (2019) also dated lateral moraines in the Kisbe Valley. The left and right lateral moraines of the Kisbe Valley (Figure 2.4a), 950 m and 800 m long respectively, and 250 m wide, are situated between 1690 and 1955 m a.s.l. The maximum thicknesses of both lateral moraines are *c.* 60 m. The authors reported four samples from the Kisbe Valley; two from the right lateral moraine (12.1 ± 1.1 ka and 5.4 ± 0.5 ka) and, two from the left lateral moraine (10.9 ± 1.1 ka and 5.6 ± 0.4 ka). As the young ages were located on the steep-sloped lower parts of the moraines, the authors claimed that the lower parts of the moraines were probably eroded and exhumed after the deposition (Köse et al., 2019). Therefore, the samples with young ages might have underestimated the true age and hence considered outliers. The mean of the two oldest ages (11.5 ± 0.8 ka) is assigned as the age of the

lateral moraines. Although Köse et al. (2019) interpret this age as probable Younger Dryas, within error margins it could be also attributed to the Early Holocene.

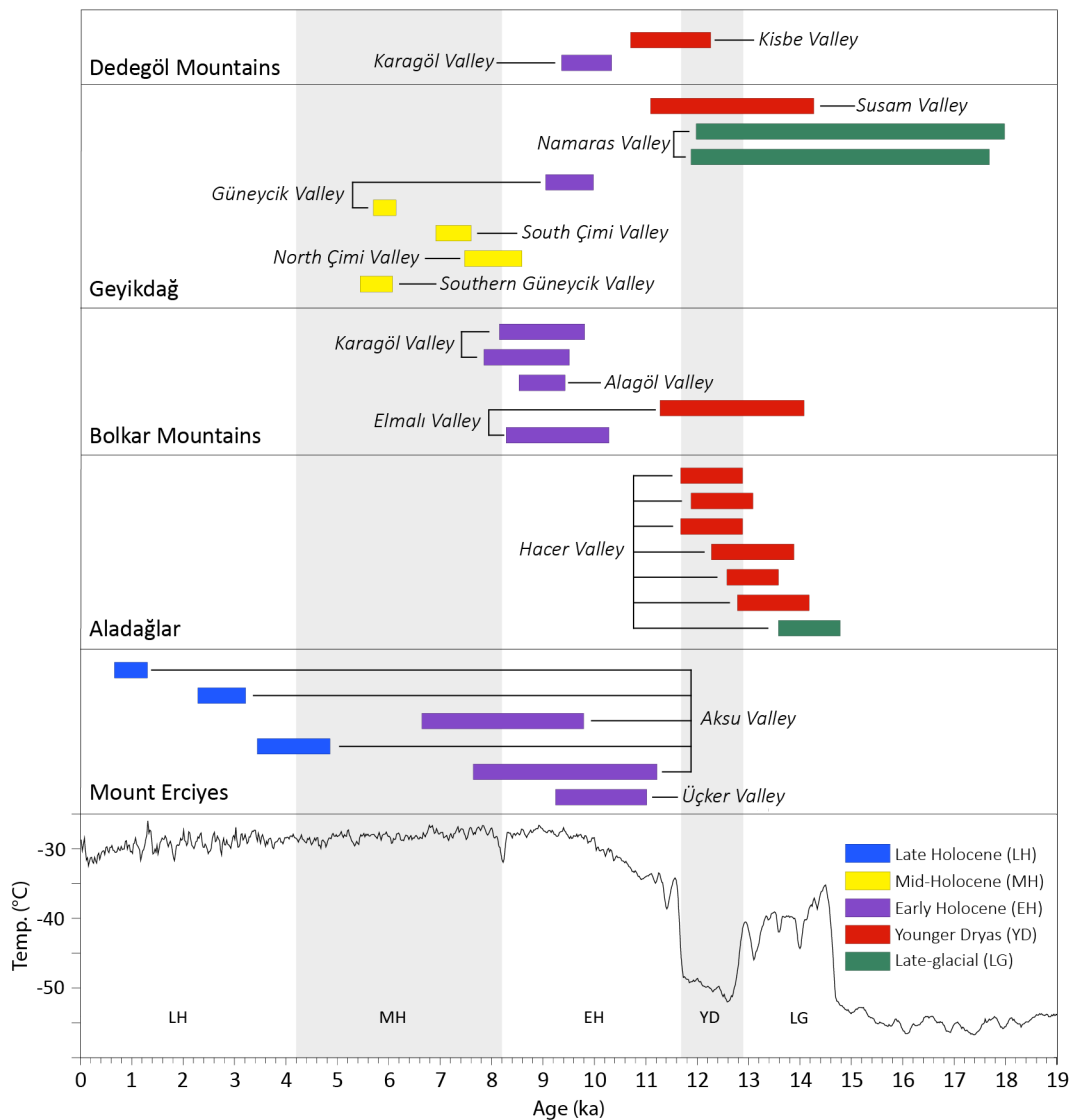


Figure 2.3 : Correlation of Lateglacial and Holocene glacial chronologies of Türkiye. Reconstructed air temperatures from the Greenland Ice Sheet Project 2 ice core (Buizert et al. 2018) are given in the lower panel. The color-coded bars reflect the assigned ages of individual moraines from original publications. These ages were recalculated based on up-to-date cosmogenic production rates and scaling schemas (Borchers et al. 2016; Marrero et al. 2016a; Phillips et al. 2016). Horizontal grey shaded/white bars are the time frames for the Late-glacial (LG), Younger Dryas (YD) stadial (11.7–12.9 ka ago; Broecker et al. 2010), Early Holocene (EH) (11.7–8.2 ka ago), Middle Holocene (MH) (8.2–4.2 ka ago) and Late Holocene (LH) (4.2 ka ago to present).

Köse et al. (2019) also collected five boulder samples from a 1.5-km-long left lateral moraine ridge on the eastern Karagöl Valley. The sample ages range from 3.7 ± 0.3 ka to 12.0 ± 1.0 ka. The average of the youngest two ages (3.8 ± 0.2 ka) is several times

younger than the average of the oldest ages (9.8 ± 1.4 ka). Therefore, they preferred to report the age of the lateral moraine as 9.8 ± 1.4 ka and claimed that the younger ages were due to the exhumation of the samples. Their results indicate Early Holocene deglaciation in the Karagöl Valley.

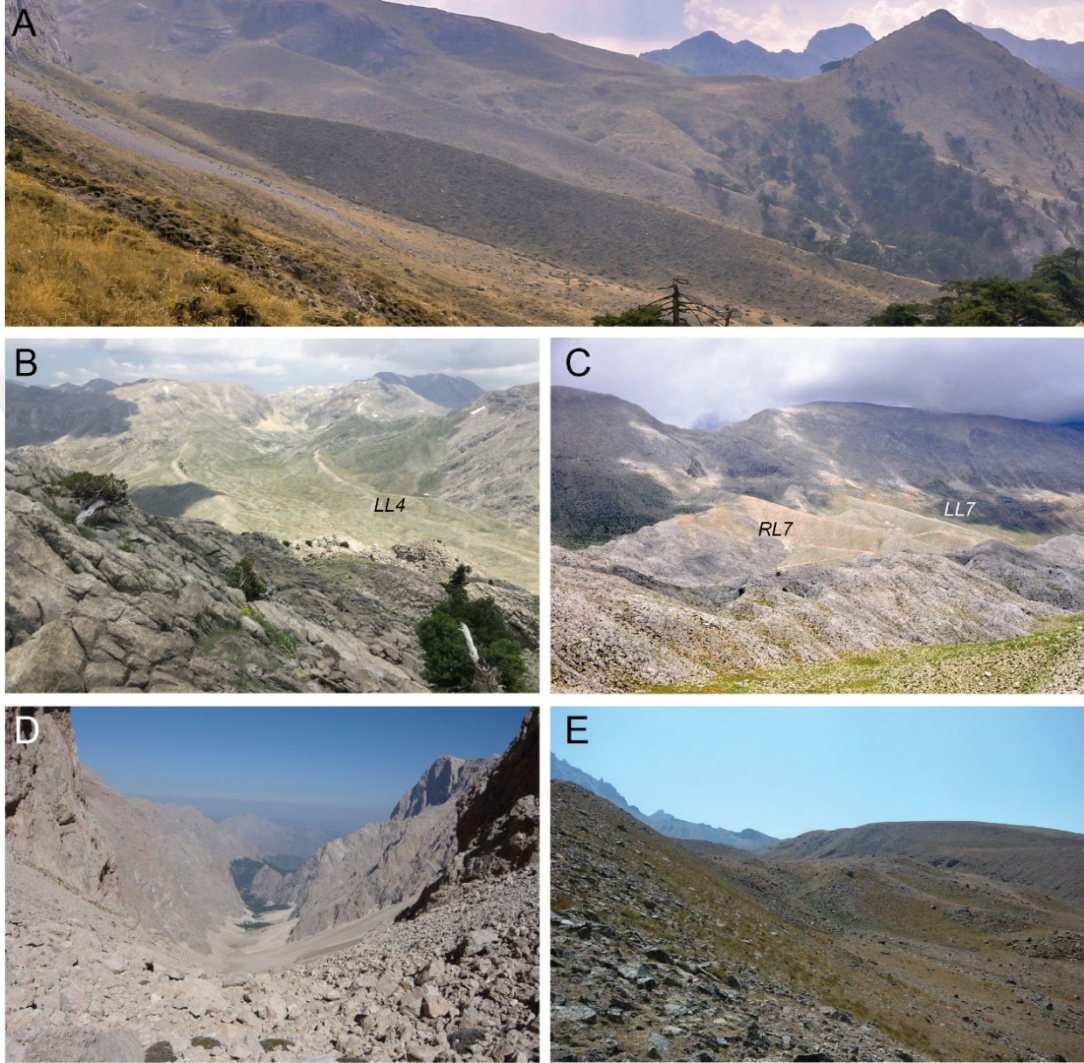


Figure 2.4 : (a) Younger Dryas lateral moraines (11.5 ± 0.8 ka) in the Kisbe Valley of Mount Dedegöl (Köse et al. 2019); (b) Middle Holocene (7.3 ± 0.6 ka) left lateral moraine (LLM4) in the south Çimi Valley of Geyikdağ (Sarıkaya et al. 2017); (c) Right (RLM7) and left (LLM7) lateral Early Holocene (9.5 ± 5.5 ka) moraines of the Güneycik Valley of Geyikdağ (Sarıkaya et al. 2017); (d) the Hacer Valley of Aladağlar, looking to the east; and (e) Early Holocene (10.2 ± 0.9 ka) moraines of Üçker Valley, Mount Erciyes (Sarıkaya et al. 2009).

OSL (Optically Stimulated Luminescence) method has been also used to date Dedegöl Mountains moraines by Çılğın (2015). However, the obtained OSL ages yielded unreliable results for glacial periods varying between the Middle Pleistocene (148 ± 13 ka) to Late Holocene (2.6 ± 0.1 ka). Zahno et al. (2009) did not report any Holocene

deglaciation ages from the Dedegöl Mountains. It is possible that the northern part of the mountain experienced glacial conditions until the onset of the Holocene (Köse et al., 2019).

2.4.2 Geyikdağ

Geyikdağ (36° 53' 22" N, 32° 10' 16" E, 2877 m a.s.l.) is situated in south-central Türkiye, *c.* 38 km north of the Mediterranean coast (Figure 2.2), between the towns of Akseki and Gündoğmuş in the province of Antalya. Because of the thrust tectonics that affected the area, the geology of the Taurus Mountains around the study area is complex (Monod, 1977; Poisson, 1977; Şenel, 1996). Although they vary in age (Cretaceous to Eocene), most of the exposed units are mainly made up of limestones (Çiner et al., 2015).

Several glacial valleys (Namaras, Susam, Çimi, Güneycik, and Çündüre valleys), with elevations ranging from 1600 to 2500 m a.s.l., contain evidence of paleoglacial events. In the Namaras Valley (2000-2050 m a.s.l.) Çiner et al. (1999) reported the largest hummocky moraine field found in the eastern Mediterranean region (*c.* 30 km²). Geyikdağ is the most important mountain that contains seven Holocene moraines compiled during this review.

The first comprehensive studies about the moraines on Geyikdağ were carried out by Çiner et al. (1999) and Çiner (2003). However, these studies were mostly related to the sedimentological description of the moraines with no absolute dating. Nevertheless, the Holocene glacial activity was indicated by the freshness of the perched cirques and arêtes, which occur at altitudes of 2600 m in the Susam Valley (Çiner et al., 1999). Some of the cirques they reported were bounded by perennial snow masses.

Later, Çiner et al. (2015) dated 34 moraine boulders in the same glacial valleys. They sampled two right lateral moraines in the Namaras Valley. The first right lateral moraine (RLM1 in Çiner et al., 2015) shows a systematic decrease in ages towards the source area (from 14.0 ± 2.7 ka to 7.9 ± 1.6 ka). They preferred to report the oldest boulder age (14.0 ± 2.7 ka) as the beginning of the deglaciation as the glacier depositing this moraine probably melted down gradually with resulting ages between 14.0 ± 2.7 ka and 7.9 ± 1.6 ka (Figure 2.3). Similarly, an 850 m long and 50 m wide, second right lateral moraine (RLM2 in Çiner et al., 2015) in the same valley sourced

from a cirque area facing north. Four boulders on that ridge have also decreasing ages towards the cirque area (from 13.6 ± 2.7 ka to 5.2 ± 1.0 ka). They also reported the oldest boulder age as the landform age (13.6 ± 2.7 ka) and claimed that the glacier depositing this moraine also probably gradually melted down between 13.6 ± 2.7 ka to 5.2 ± 1.0 ka (Çiner et al. 2015). Their results indicate a gradual retreat from Late-glacial to mid-Holocene, and this time interval incorporates the Younger Dryas (*c.* 12 ka) period in the Geyikdağ. Three samples from the third right lateral moraine (RLM3 in Çiner et al., 2015) at the northern end of the Susam Valley yielded surface exposure ages of 10.3 ± 2.1 ka, 12.2 ± 2.3 ka and 12.3 ± 2.4 ka (weighted average age of 11.6 ± 1.3 ka) indicating the Younger Dryas to Holocene deglaciation in this nearby valley.

Sarıkaya et al. (2017) reported a comprehensive Holocene glacial phase in the western and southern valleys of the Geyikdağ. The last glaciation in the mountain occurred during the Holocene and ended between 9.6 ± 1.4 ka and 5.9 ± 0.5 ka ago. Later, glaciers mostly vanished from the Geyikdağ, but a few rock glaciers developed during the Late Holocene (Sarıkaya et al., 2017). In the west-facing Çimi Valley, the authors collected four samples from the left lateral moraine (Figure 2.4b) (LLM4 in Sarıkaya et al., 2017) ranging from 6.6 ± 0.6 ka to 8.7 ± 0.8 ka that yielded an average age of 7.3 ± 0.6 ka, which indicates a Holocene glacial phase. On the northern Çimi Valley, they also collected two samples (8.6 ± 0.8 ka and 7.5 ± 0.6 ka) from the LLM5 moraine that yielded an average age of 8.1 ± 0.9 ka, indicating synchronous deglaciation during the Holocene.

The authors also collected six samples that range from 4.6 ± 0.4 ka to 10.5 ± 1.0 ka from the east-facing 2.5 km-long left lateral moraine (LLM6 in Sarıkaya et al., 2017) ridge in the Güneycik Valley. The average age of LLM6 is calculated as 5.9 ± 0.5 ka indicating a Middle Holocene activity, which is also comparable with the Holocene moraines of the Çimi Valley. Besides, two well-preserved moraine ridges (Figure 2.4c) (RLM7 and LLM7 in Sarıkaya et al., 2017) are located on the south-facing part of the Güneycik Valley near Güneyçukuru. Three samples that range from 7.5 ± 0.6 ka to 11.2 ± 1.0 ka on the RLM7 moraine yielded an arithmetic average age of 9.6 ± 1.4 ka.

Holocene moraines in the southern (9.5 ± 0.5 ka) and eastern (5.9 ± 0.2 ka) slopes of the Güneycik Valley along with those formed in the north-facing Namaras Valley (6.0 ± 1.0 ka) (Çiner et al., 2015) and the western Çimi Valley (7.3 ± 0.6 ka and 8.1 ± 0.9 ka) are perhaps some of the best examples of Holocene glaciations in the Taurus

Mountains. It is also interesting to note that the south-facing Holocene glacier appears to have retreated several thousand years before the glaciers in the northern direction.

Last but not least, Çiner et al. (2017) dated two different geomorphic units in southern Güneycik Valley; a well-preserved moraine loop (the so-called Zor Moraine) and a rock glacier inside that loop (Figure 2.5), which has contrasting morpho-stratigraphic positions and cosmogenic ^{36}Cl ages. They obtained a Middle Holocene age (5.8 ± 0.3 ka) from the moraine loop with a weighted average of four measured samples (one of these samples is an amalgamated sample; see Sarıkaya et al., 2016).

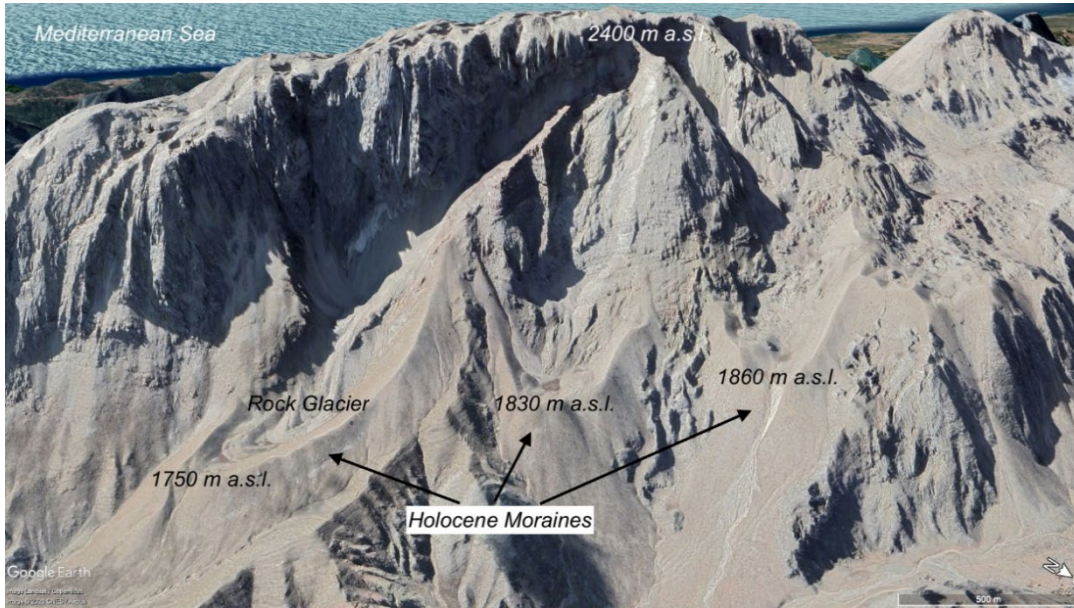


Figure 2.5 : Google Earth view of the Güneycik Valley's north-facing slopes of Geyikdağ.

2.4.3 Bolkar Mountains

The Bolkar Mountains ($37^{\circ} 23' 24''$ N $34^{\circ} 36' 36''$ E, 3524 m a.s.l.) are situated in south-central Türkiye, 65 km north of the Mediterranean coast (Figure 2.2). The mountain range is broadly parallel to the sea and aligned in a NE-SW direction. It rises from sea level through an area of 90 km long and 50 km wide to elevations between 2000 and 3500 m a.s.l. Paleozoic and Mesozoic green schists, marbles, limestones, and interbedded plutonics form the bulk of the rocks in the central mountain range of the Taurus (Louis, 1944; Blumenthal, 1956; Demirtaşlı, 1984). Limestones dominate as the bedrock in the study area, but also as till with several other erratic boulders of plutonic origin (gabbro, diorite, etc.) transported by glaciers (Birman, 1968).

Although several valleys were glaciated in the past in the Bolkar Mountains, Çiner & Sarıkaya (2017) have chosen to study the most representative ones. The Karagöl, Alagöl, and Elmalı valleys were studied via cosmogenic ^{36}Cl surface exposure dating. Glaciers in the Karagöl and Alagöl valleys advanced from well-defined cirque areas to the north and northeast, reaching down to about 2200 m a.s.l.

Towards the Meydan Meadow on the northeastern Karagöl Valley, two lateral moraines start at 2450 m a.s.l., descend to 2200 m a.s.l., and attain 1.5 km in length. Çiner & Sarıkaya (2017) sampled two boulders from the right lateral moraine (RLM2) that yield similar ages (8.0 ± 1.3 ka and 10.4 ± 1.7 ka) representing an Early Holocene retreat. From the counterpart left lateral moraine (LLM2), their ages gave a weighted average of 9.0 ± 0.9 ka, which also indicates an Early Holocene retreat of the paleoglacier from the Karagöl Valley (Figure 2.3).

Four Early Holocene ages were obtained in three different valleys (Karagöl, 9.0 ± 0.8 ka, 8.7 ± 0.8 ka; Alagöl, 9.0 ± 0.4 ka, and Elmalı, 9.3 ± 1.0 ka) (Çiner & Sarıkaya, 2017). The Late-glacial moraines were dated to between 16.8 ± 0.7 and 12.7 ± 1.4 ka in both northern Alagöl and southern Elmalı valleys, respectively. In the Early Holocene, the final deglaciation occurred in the Karagöl and Elmalı valleys respectively 9.1 ± 0.5 ka and 9.4 ± 0.9 ka ago. Contemporaneous deglaciation (9.4 ± 0.9 ka) was also observed in the south-facing Elmalı Valley and the outwash deposits (9.0 ± 0.4 ka) of Alagöl Valley (Çiner & Sarıkaya, 2017). According to Çalışkan et al. (2014), seven debris-covered glaciers, the largest with an area of 0.23 km^2 , still exist on the mountain.

2.4.4 Aladağlar

Aladağlar ($37^\circ 47' 48'' \text{ N}$, $35^\circ 9' 20'' \text{ E}$, 3756 m a.s.l.) shows evidence of previous glaciers (Klaer, 1962; Klimchouk et al., 2006) and contain a whole range of ages from pre-LGM to Holocene (Köse et al., 2018). The mountain range consists of extensively karstic Mesozoic carbonates that restrict surface drainage (Tekeli et al., 1984; Klimchouk et al., 2006).

Zreda et al. (2011) collected and analyzed 22 samples from seven lateral and terminal moraines in the Hacer Valley (Figure 2.4d), the longest and deepest eastern-facing valley of the mountain range. Twenty of these samples were collected from boulders and two from glacially scoured bedrocks, between 1100 and 3100 m a.s.l.

Sarıkaya & Çiner (2017) recalculated these 22 samples using new cosmogenic production rates given by Marrero et al. (2016a). The ages indicate the start of deglaciation at the lower limit of the valley at 1100 m in Late-glacial (14.0 ± 1.5 ka) (Figure 2.3). The *c.* 15 km long glacier rapidly melted and at the onset of the Holocene (11.5 ± 1.3 ka ago) the high plateau (Yedigöller Plateau) was mostly ice-free. The authors also concluded that deglaciation occurred in two phases and the second phase was faster than the first one. In conclusion, they indicate that the glaciation in the Hacer Valley not only was unusually extensive for its age but also the deglaciation rates and the inferred climate changes were faster than modern rates.

Other glacial valleys in the Aladağlar indicate glaciers from much older periods (Köse et al., 2018). In the west and northwest sections of the Aladağlar, there are several valleys with terminal and lateral moraines that predate global LGM. These contain evidence that glaciation also developed in MIS 3 or MIS 4 periods based on cosmogenic surface exposure ages obtained from western-facing outwash deposits that were later cut and displaced by the Ecemiş Fault Zone (Sarıkaya et al., 2015a, b). Few small LGM moraines (21.9 ± 1.8 ka) were reported from the western Aladağlar that was later cut by the mountain front Cevizlik Fault (Yıldırım et al., 2016).

2.4.5 Mount Erciyes

Mt. Erciyes ($38^{\circ} 31' 55''$ N, $35^{\circ} 26' 56''$ E; 3917 m) (Figure 2.2) is a stratovolcano in Cappadocia Volcanic Province in central Anatolia (Şen et al., 2003). The youngest volcanic deposits were reported from Perikartını, Karagüllü, and Dikkartın rhyodacite domes, yielding ^{36}Cl and ^{14}C ages of *c.* 10 ka (Sarıkaya et al., 2019).

Mt. Erciyes is the only mountain where Late Holocene glacial deposits were reported in Türkiye. Since the LGM, and when the glaciers were the most extensive, reaching 6 km in length and descending to an altitude of 2150 m a.s.l., three retreats and advances occurred in Mt. Erciyes (Sarıkaya et al., 2009). The first retreat started in LGM. Two readvances and retreats occurred in series in the Late-glacial and the Early Holocene. Finally, the latest advance took place in the Late Holocene (Sarıkaya et al., 2009). Today, a small glacier (*c.* 260 m long) is present to the north of the peak, with a snowline elevation of 3550 m (Sarıkaya et al., 2011).

In the upper part of a tributary valley, Sarıkaya et al. (2009) obtained three samples from fresh-looking moraine boulders in Aksu Valley. Their mean age (4.1 ± 0.7 ka)

suggests that these were deposited during the Middle Holocene (Figure 2.3). Three additional samples were collected from moraines in Aksu Valley with a mean age of 1.0 ± 0.3 (outlier excluded). The ages of the two tributary cirques/valleys indicate that the upper moraines in the Aksu Valley were deposited during the Late Holocene (Sarıkaya et al., 2009).

The lower outwash plains, between 2600 and 2700 m a.s.l., contain numerous large boulders scattered away from the ridges of the bedrock (Sarıkaya et al., 2009). The authors gathered five samples from the boulders on those outwash plains to support their moraine ages from the upper Aksu Valley. The samples yielded ages between 2.2 ± 0.7 ka and 11.0 ± 2.8 ka. The ages were split into two groups based on age and distance within the outwash plain; the first group with an average age of 2.8 ± 0.5 ka and the second group with an average age of 8.3 ± 1.6 ka. The first group suggests that they were deposited in the upper valley during the younger glacier advances (Late Holocene) and the second group during the older advances (Early Holocene) (Sarıkaya et al., 2009).

In the Üçker Valley, Sarıkaya et al. (2009) collected eight samples from the hummocky moraine complex between 2650 and 3100 m (Figure 2.4e). Their ages range from 8.0 ± 2.0 ka to 13.0 ± 3.0 ka with an average age of 10.2 ± 0.9 ka, which suggests they were deposited during the glacial advances of the Early Holocene. This is consistent with the Early Holocene moraines in the Aksu Valley, dated to 9.5 ± 1.8 ka. Therefore, the average age (9.3 ± 0.9 ka) of moraines in both valleys was proposed as the time of deglaciation that translates to Early Holocene.

2.5 Recent Glaciers in Türkiye

Although Holocene glacial deposits are rather located in the west and central Taurus Mountain Range, active glaciers are mostly located in the east, northeast, and southeast Anatolia. Thirty-three glacierets, 17 mountain glaciers, 1 ice cap (51 features in total), and 55 rock glaciers were recognized by remote sensing methods (Sarıkaya & Tekeli 2014). According to Sarıkaya & Çiner (2019), the ice cap of Mt. Ağrı alone constitutes half of the total glaciated area of Türkiye (5.66 km^2 of 11.5 km^2 in total). In addition to this, the longest and most noticeable mountain glaciers are located in the southeastern Taurus Range, on Mt. Buzul. İzbirak and Erinç glaciers are 2.1 km and 1.5 km in length, respectively, both covering an area of about 2 km^2 . There are many

smaller glaciers, smaller than 0.2 km², which are probably remnants from earlier glacial stages (Sarıkaya & Çiner, 2019). Mt. Erciyes as a volcanic mountain just like Mt. Ağrı also has a small glacier (c. 260 m long with an area of 0.05 km²) that is present to the north of the peak, with a snowline elevation at 3550 m a.s.l. (Sarıkaya et al., 2011). Although the existence of other glaciers in the eastern Black Sea Mountains and Taurus Mountain Ranges are also known, they are only a few hundred meters long and most of them are covered by debris (Sarıkaya & Çiner, 2015).

The geomorphology of Mt. Ağrı was studied by Azzoni et al. (2017), who focused on glacial and periglacial, water-related, gravitational, and volcanic landforms. Azzoni et al. (2017) provide Mt. Ağrı's first geomorphological map and record the first detailed quantification of the active glacier, based on field and satellite data. They also provide the first overview of three, previously unknown, debris-covered glaciers located in this area, covering 25% of the total surface of the ice. On the other hand, Azzoni et al. (2020) identified 51 ice bodies covering 12.29 km², in their most recent study on Turkish glaciers. The authors indicated that the Mt. Ağrı ice cap and four small outlet glaciers cover 7.37 km², which represents >60% of the Turkish glacierized area. In a very recent study, Yalcin (2020) mapped glacier vulnerability areas of Mt. Ağrı using GIS-based Multi-Criteria Decision Analysis and reported a strong relationship between glacier surface temperature anomalies, direction, and elevation in calculating glacier retreat.

Azzoni et al. (2020) grouped active Turkish glaciers into 5 areas: in addition to the main ice cap on Mt. Ağrı. There are 12 glaciers in the Munzur Mountains (0.31 km²), 29 glaciers in the southeast Taurus Mountains (Buzul Mountains) (3.82 km²), 3 ice bodies in Mt. Süphan (0.20 km²), and six glaciers in the eastern Black Sea Mountains (Kaçkar Mountains) (0.58 km²). They did not consider the glacier on the slope of Mt. Erciyes, defined by Sarıkaya et al. (2009), as its area is now below 0.01 km², a standard threshold used in regional glacier inventories (Paul et al., 2009).

Following the warming at the onset of the Holocene, mountainous areas were subject to paraglacial and periglacial activities (see Oliva et al., 2018 for a review). As a result, rock glaciers developed in several mountains like Mt. Erciyes (Sarıkaya et al., 2009), Mt. Kaçkar (Akçar et al., 2007), Karçal Mountains (Dede et al., 2017), Geyikdağ (Çiner et al., 2017) and Mt. Karanfil (Çiner et al., 2018; Köse et al., 2018). Although some of these rock glaciers were dated to Late-glacial (c. 16 ka), in Karçal Mountains

(Dede et al., 2017) and on Mt. Karanfil they indicate the onset of Holocene (*c.* 10 ka) (Çiner et al., 2018).

2.6 Discussion and Conclusion

2.6.1 Mediterranean Holocene glacial records

Apart from cosmogenic dating of glacier-related landforms (moraines, erratic boulders, and glacier-sculptured bedrock) several other proxies such as ^{14}C , OSL, lichenometry, ice core, tree rings, speleothems, and varves are used in the reconstructions of glacier fluctuations. A wealth of information is available in the literature related to Holocene glacial and paleoclimate change and the readers are referred to Solomina et al. (2015) for a comprehensive review of the Holocene glacier fluctuations around the globe. On the other hand data on the Holocene glacial record of the Mediterranean region are still relatively scarce compared to the LGM record but numerous works have been published in special issues (e.g., Davis & Osborn, 1988; Davis et al., 2009; Oliva et al., 2020a, b).

It appears that during the Early Holocene, glacier recession was underway in most areas of the Northern Hemisphere, including the Mediterranean, with some glaciers completely vanishing (Davis et al., 2009). For instance, climatic conditions in the Alps did not permit glacier advances between *c.* 10.5 ka and 3.3 ka (Ivy-Ochs et al., 2009). Nevertheless, during a few short intervals, such as a cooling event *c.* 8.2 ka ago, small glaciers ($<0.5 \text{ km}^2$) -81% of glaciers in the Alps are small and therefore sensitive to climate change (Zemp et al., 2006)- might have advanced close to their Little Ice Age (LIA) extents, but these cold intervals did not last long enough for large glaciers to advance (Solomina et al., 2015, 2016). In the French and Spanish Pyrenees glacier fluctuations during the Holocene are also extensively documented (Pallàs et al., 2006; Palacios et al., 2017; Jomelli et al., 2020).

On the onset of Neoglaciation, which corresponds to general cooling periods starting *c.* 5 ka in the North Atlantic, phases of glacial advances were also recognized in Italian Apennines (4.3 and 1.3 ka; Giruadi, 2004), in the Maritime Alps (LGM to LIA; Federici et al., 2017), in eastern Switzerland (between 3.3 and 2.8 ka; Schimmelpfennig et al., 2014), in Spanish Sierra Nevada (2.8-2.7 cal ka BP; Oliva et

al., 2020a, b), and high cirques of Mt. Olympus in Greece (between 15.5 and 0.64 ka; Smith et al., 1997; Styllas et al., 2018).

LIA cold period (c. 1270-1860 AD) witnessed some of the most important Holocene glacier advances in the Northern Hemisphere (Davis et al., 2009). On the other hand, satellite data show that the Swiss glacier area loss from 1985 to 1999 increased seven-fold compared to the period from 1850 to 1973 (Paul et al., 2007). This probably indicates the anthropogenic effect on the retreat of glaciers and further research will be necessary to understand the extent of this deglaciation on a global scale.

2.6.2 Turkish Holocene glacial record

Recent cosmogenic surface exposure dating studies in Türkiye have increased our current understanding of the extent and timing of glaciations in the Anatolian mountains. However, most of the ages indicate extensive LGM and Late-glacial glaciations but Holocene ages are rather scarce. Holocene ages were only reported from western and central Taurus Range and central Anatolia from 5 mountains scattered along 14 valleys as explained above. On the other hand, no Holocene glacial activity was reported from the western Taurus (i.e. Mt. Sandıras, Akdağ), the northeastern Black Sea Mountains (i.e. Kaçkar Mountains, Akçar et al., 2014, 2015) or Uludağ on northwest Türkiye (Zahno et al., 2010).

The elevation of the southwestern Taurus Mountains is generally less than 3000 m a.s.l. For instance, Mt. Sandıras is the lowermost and westernmost glaciated mountain in Türkiye and has a peak elevation of 2295 m a.s.l. There were only LGM advances (22.9 ± 3.3 ka and 20.6 ± 3.1 ka) and later, minor re-advances or stabilizations during Late-glacial (16.4 ± 3.2 ka and 17.2 ± 3.2 ka) (Sarıkaya et al., 2008). The LGM time ELA (c. 2000 m a.s.l.) was the only c. 300 m below the summit altitude. Therefore during the Holocene, the Holocene-ELA was probably not below the peak altitude, and hence it did not allow accumulation of ice on the mountain. This may be also true for other southwestern Taurus Mountains, i.e. Akdağ (Sarıkaya et al., 2014), even though this mountain has a higher peak altitude (3016 m a.s.l.). Although the northeast Black Sea Mountains have much higher altitudes (3900-4000 m a.s.l.) and therefore might sustain Holocene glaciations, they have not been mapped and/or dated yet.

After the LGM, glaciers in the Anatolian mountains became smaller during the Late-glacial and Younger Dryas. There is no clear evidence of re-advances of the existing

LGM glaciers during the Late-glacial. Although during the Younger Dryas there were some advances (e.g., Hacer Valley of Aladağlar), the onset of the Holocene is mostly related to glacier retreats as in other parts of the world.

Five study areas presented in this review show that the Holocene glaciations were concentrated on the western and central Taurus Mountain Range and Mt. Erciyes, in central Anatolia. The Late-glacial moraines in the central Taurus Mountains show widespread distribution from 14.8 ± 2.9 ka to the onset of the Holocene. During the Holocene, these mountains also hosted Early Holocene (11.5 ± 0.8 ka) and Middle Holocene (5.8 ± 0.3 ka) glaciers.

On the other hand, Mt. Erciyes, the only mountain with a Late Holocene chronology, has no traces of the Middle Holocene period. According to Sarıkaya et al. (2009), the Late Holocene moraines on Mt. Erciyes show that glaciers advanced and retreated 4.1 ± 0.7 ka ago, and again in the Aksu Valley 1.0 ± 0.3 ka ago. The recent glacier on Mt. Erciyes was interpreted as the remnant of the Late Holocene glacial advance by Erinç (1952), with terminal moraines located along the front of the retreating glacier reflecting phases of glacial regeneration and expansion in a cooler and wetter climate than today (Sarıkaya et al., 2009).

LIA moraines appear to be absent in many Turkish mountains. Several active and passive rock glaciers that developed in higher altitudes obscure the relicts of glaciers, and also most likely, the LIA moraines in the uppermost parts of the valley system (Akçar et al., 2008). However, the main valley in the Aksu Valley of Mt. Erciyes (above 3000 m a.s.l.) has very fresh-looking moraines, which might be related to the LIA (Sarıkaya et al., 2009). The modern glacier on Mt. Erciyes is interpreted as a remnant of the LIA glacier due to the similar characteristics of its deposits (Erinç, 1952). LIA glacier remnants in other Turkish Mountains remain to be studied in detail.

Today the easternmost part of the Taurus Range, the northeast part of the Black Sea Mountains, and some other individual mountains (e.g., Ağrı, Süphan, and Munzur mountains; Çılğın, 2015; Bayrakdar et al., 2015) host glaciers (Azzoni et al., 2020). Nevertheless, Holocene glacial chronologies are still missing from the eastern part of Türkiye. Considering the extent of paleoglacial/glacial regions, further studies should focus on the undated mountains that would substantially increase our understanding of the glacial chronologies and paleoclimate of Türkiye.



3. COSMOGENIC ^{36}Cl SURFACE EXPOSURE DATING OF GLACIAL LANDFORMS ON MT. BARLA (SW TÜRKIYE)²

3.1 Introduction

Postglacial processes often modify landforms developed by glaciers during interglacial periods (Evans & Clague, 1988, 1994). Because of the severity of these processes, identifying the paleoglacial environments and applying numerical methods such as cosmogenic surface exposure dating to understand the glacial history might be problematic. Based on the potential effect of postglacial processes on paleoglacial landscapes, it may be challenging to distinguish between glacial landforms. At this point, lithology plays a crucial role, as the erosion rate of the glaciers and the preservation of the glacial-derived erosional-depositional landforms may depend on it. In carbonate paleoglacial landscapes, it is possible to see the effect of both physical and chemical weathering as a cause of deformation (e.g., doline formation in moraines, freeze-thaw-based rockfalls/rock avalanches on cirque headwalls). Changes due to postglacial karstification may have a negative impact on deglaciaded topography in the context of feature identification (Alonso, 1998). These modifications were defined as ‘karstiglacial’ landforms, the glacial origin features adapted to karstic drainage (Ford, 1979).

On the other hand, the carbonate rock-dominated lithologies may preserve the paleoglacial topography, as there is no significant surface water that would reshape the topography by fluvial action (Woodward et al., 2004; Hughes et al., 2006; Žebre & Stepišnik, 2015; Çiner et al., 2017). In contrast to karstiglacial processes, the term ‘glaciokarst’ (also known as ‘glaciated karst’) refers to landscapes characterized by karst features on inherited erosional structures from previous glaciations (Smart, 2004). For instance, preglacial karstic depressions may have been transformed into

² This chapter is based on the paper “Altınay, O., Sarıkaya, M. A., Çiner, A., Žebre, M., Stepišnik, U., Yıldırım, C., Yetemen, Ö., & Wilcken, K. M. (2022). Cosmogenic ^{36}Cl surface exposure dating of glacial landforms on Mt. Barla (SW Turkey). *Geomorphology*, 416, 108424.”

cirques by glacial erosion, snow erosion, and mass movements (Fels, 1929; Smart, 1987).

Mt. Barla is located in the Central Taurus Mountains Karst Area, the subregion of the Taurus Mountains Karst Region (i.e., the Mediterranean Karst zone) (Nazik & Tuncer, 2010). This sub-region covers a vast area with lateral and vertical karst developed by polygenetic-polycyclic processes (Nazik et al., 2019). Furthermore, glaciokarst is a common phenomenon in this region (e.g., Bayrakdar, 2012; Çılğın, 2012; Bayrakdar et al., 2017, 2020; Çılğın & Bayrakdar, 2020). Evans et al. (2021) also remarked on the role of karstification in the glacial spatial distribution of cirques and development in the Western Taurus.

Numerous researchers have studied the glacial geomorphology and chronology of the southwestern Taurus Mountain Range (e.g., Sarıkaya et al., 2008, 2014, 2017; Zahno et al., 2009; Bayrakdar, 2012; Çılğın, 2012; Çiner et al., 2015; Çiner & Sarıkaya, 2017; Köse et al., 2019; Candaş, et al., 2020). Several syntheses related to glacial chronologies in the Taurus Range and Türkiye in general are also available (e.g., Akçar & Schlüchter, 2005; Sarıkaya & Çiner, 2015, 2017; Sarıkaya et al., 2011, 2017; Altınay et al., 2020; Allard et al., 2021; Akçar, 2022a, b, c). Oliva et al. (2018) also provide a general outline of periglacial environments in Türkiye. Nevertheless, a proper glacial geomorphological map and glacial chronology are crucially lacking on Mt. Barla. Therefore, the primary objectives of this study are to 1) provide comprehensive geomorphological maps of the study area, allowing easier recognition of the geomorphological features and the polygenetic-polycyclic evolution, and 2) reveal the cosmogenic ^{36}Cl surface exposure geochronology of Mt. Barla by focusing on the glacial landforms developed during the Late Pleistocene.

3.2 Study Area

Mt. Barla (also known as Mt. Gelincik, 38° 3' 10" N, 30° 42' 7" E, 2799 m a.s.l., above sea level) is located in the Taurus Range of SW Türkiye (Figure 3.1a, b). This area is considered the eastern boundary of the Western Taurus Mountains and forms the transition from the coastal Taurus to the interior Anatolia Plateau (Nazik & Tuncer, 2010; Evans et al., 2021). To the east, the mountain is bordered by Eğirdir Lake (916 m a.s.l.) and alluvial plains to the north and south.

The Taurus Mountain Range extends parallel to the Mediterranean coast of Türkiye. It is characterized by high peaks (>3000 m a.s.l.), reshaped by glaciers during the Late Pleistocene. There are three main glacial valleys on Mt. Barla. The most prominent one, Gedik Valley, is located on the northeast side of the mountain. Kapıdere and Ortayazı valleys are less developed and situated north and northwest of Mt. Barla, respectively (Figure 3.1c).



Figure 3.1 : (a) Mt. Barla location map; (b) the boundary of the study area is shown with the green rectangle. (1) Barla and (2) Eğirdir meteorology stations; (c) cirques and valleys of the study area shown on the Red Relief Image Map (RRIM) (GC: Gelincik Cirque; GaC: Gelincikana Cirque; İC: İkizler Cirque; TC: Toklu Cirque; and AC: Akdere Cirque).

The Taurus Mountain Range was separated from Gondwana in the Triassic by the southern branch of the Neo-Tethys Ocean (Okay, 2008; Şengör et al., 2019). The Western Taurus Mountains consist of carbonate platforms (autochthonous and parautochthonous units) overlain by terrestrial sediments and nappe complexes of ophiolites. Mt. Barla is part of this carbonate platform and resides in the Barla Unit (Gutnic, 1971; Yalçınkaya, 1989). Our study area comprises Upper Anisian-Norian dolomites, Liassic algal limestones, and Dogger-Cenomanian cherty limestones and

calciturbidites (Balçı, 2011). The paleoglaciers in the mountain developed substantially on the algal/cherty limestones.

The Barla Meteorology Station (in Barla Village, 1187 m a.s.l.) (Figure 3.1b) is the closest station to the study area, located at the southeastern foot of Mt. Barla, within the boundaries of the Eğirdir Lake Basin. According to the short-term mean annual data of the station (2018-2019), the MAT (mean annual temperature) is 13.8 °C, and the MAP (mean annual precipitation) is 542 mm. Moreover, the long-term data (years 1930-2020) of Eğirdir Meteorology Station (in the Eğirdir city center, 920 m a.s.l.) (Figure 3.1b) show that the MAT is 12.7 °C (with 59 years of observations). The MAP data of the Eğirdir Meteorology Station were measured manually (years 1930-2004) and automatically (2005-2020) using AWOS (automatic weather observation station). While the manual measurements yielded 781 mm for 60 years of observation, the AWOS measurements provided 816 mm for the Eğirdir city center. The temperature lapse rate (0.0065 °C/m; Kint et al., 2014) shows that the MAT at the peak of Mt. Barla (2799 m a.s.l.) is 0.5 °C.

Research about Pleistocene glaciations on Mt. Barla is limited to the studies of Ardos (1977a), who reported multiphase glaciations (at least two glaciations) during the Late Pleistocene. Recently, Evans et al. (2021) also investigated the morphometric features of the cirques on Mt. Barla and indicated that the ELA was 2450 m a.s.l. during Pleistocene glaciations.

3.3 Material and Methods

3.3.1 Fieldwork and geomorphological mapping

In 2019 and 2020, we conducted two fieldwork campaigns to map the mountain and collect cosmogenic surface exposure dating samples. We collected 11 moraine block samples in the Gedik and Kapıdere valleys during the first fieldwork (Figure 3.1c). The second fieldwork was carried out to acquire oblique images by drones to produce detailed geomorphology maps. The maps were created using satellite imagery from Google Earth, Küre-HGM (HGM: The General Directorate of Mapping of Türkiye; *Tr. Harita Genel Müdürlüğü*), DigitalGlobe (Maxar), digital surface models (DSM) from HGM and oblique imagery captured by DJI Phantom 3 Professional and DJI Mavic Pro drones in the field. We also used maps produced by Ardos (1977a) to map

the paleoglacial morphology. In addition, we applied the Red Relief Image Map (RRIM) visualization technique with digital surface model data to map geomorphic features (Figure 3.1c). The RRIM approach is based on blending three topographic element layers: slope, positive openness, and negative openness (Chiba et al., 2008). SAGA GIS (v. 7.9.0) was used to create positive and negative openness layers (Yokoyama et al., 2002). While valleys and depressions have higher negative openness values, ridges have higher positive openness values.

3.3.2 Cosmic ray exposure dating

We used cosmogenic surface exposure dating to understand the retreat of paleoglaciers on Mt. Barla. Cosmogenic ^{36}Cl can be applied to carbonate rocks, the most widespread lithology in our study area. This technique was previously successfully applied to several other sites in Taurus Mountains Range (Zreda et al., 2011; Sarıkaya et al., 2008; Sarıkaya et al., 2014; Çiner et al., 2015; Çiner and Sarıkaya, 2017; Çiner et al., 2017; Köse et al., 2019, 2022).

The uppermost layers of moraine boulders (>0.3 m in height) on stable or gently sloping surfaces were sampled using a hammer and chisel. Eleven carbonate rock samples with thicknesses between 0.8 and 3 cm were collected (Table 3.1). To minimize the chances of overturning, stable and well-embedded boulders were selected far from the influence of slope processes. To ensure optimal exposure to the cosmic rays, flat and gentle rock faces were preferred to sharp ridges and steep sides. A compass and a clinometer were used to measure topographic shielding angles at each sampling site at every 45° azimuth interval.

Table 3.1 : Field attributes of the samples (L: length, W: width, and H: height of the boulders).

	GPS Latitude (WGS84)	GPS Longitude (WGS84)	Elevation from the topographic map	Boulder dimensions (LxWxH)	Sampled thickness	Topography shielding factor
Sample ID	°N (DD)	°E (DD)	(m a.s.l.)	(m)	(cm)	(-)
BA19-01	38.06724	30.71226	2225	1.1x1x0.5	0.8	0.9487
BA19-02	38.06801	30.71553	2208	2.5x0.8x2.6	3	0.9584
BA19-03	38.06878	30.71740	2173	1.8x0.85x0.3	2	0.9571
BA19-04	38.07087	30.72080	2102	1x2.4x0.65	2	0.9628
BA19-05	38.07857	30.73519	1853	1.1x1.4x1.7	3	0.9591
BA19-06	38.07828	30.73532	1850	2.1x1x0.9	2.5	0.9564
BA19-07	38.08494	30.70756	1881	1.4x1.1x0.8	2.5	0.9517
BA19-08	38.08595	30.70708	1881	1.1x0.7x0.6	2	0.9592
BA19-09	38.08588	30.70555	1883	1.8x1x0.6	2	0.9556
BA19-10	38.08618	30.70528	1878	5.5x2x1.7	3	0.9592
BA19-11	38.08619	30.70564	1875	1.9x1.1x0.9	2	0.9574

Sample preparation was carried out at the ITU/Kozmo-Lab (Istanbul Technical University, Türkiye) (<http://www.kozmo-lab.itu.edu.tr/en>) according to the procedures described in Sarıkaya (2009) and Schimmelpfennig (2009). Rock samples were first crushed and ground to the size fraction of 0.25 to 1 mm. After leaching samples with milli-Q and dilute nitric acid (10%) to remove any meteoric chlorine, natural chlorine was liberated from the rock matrix by dissolving the sample with 400 ml 65% HNO₃ in HDPE bottles. A mixture of natural NaCl (Merc Emsure) and ³⁵Cl (~99.7%) enriched carrier (Aldrich) was added to the samples before total dissolution. Chlorine was precipitated by the addition of AgNO₃. ³⁶S is an isobar of ³⁶Cl and was removed from the samples by repeated precipitation as BaSO₄ (Mechernich et al., 2019). The chlorine isotopic ratios of the samples were measured with a 6 MV SIRIUS Tandem Accelerator, Australian Nuclear Science and Technology Organization (ANSTO), Sydney, Australia (Wilcken et al., 2017, 2019).

The primary and secondary AMS standards are Purdue Z93-0005 (³⁶Cl/Cl: 1.2 ± 10^{-12} with natural ³⁷Cl/³⁵Cl ratio) and KN supplied by Dr. Kunihiro Nishiizumi (1.6 ± 10^{-12} and 5.0 ± 10^{-13} , ³⁶Cl/Cl ratios; Sharma et al., 1990), respectively (Wilcken et al., 2013). The decay constant of $2.303 \pm 0.016 \times 10^{-6} \text{ yr}^{-1}$ was used, corresponding to a ³⁶Cl half-life of 3.014×10^5 years. The analytical uncertainties include counting statistics, machine stability, and blank correction. Total Cl was determined by the isotope dilution method (Wilcken et al., 2013). The natural chlorine concentrations are low,

ranging from 2 to 15 ppm. Major and selected trace element concentrations of all samples were measured at Activation Laboratories, Inc., Ontario, Canada (Table 3.2). CaO concentrations are 55.31% on average, while K₂O concentrations are very low, near the detection limits (0.01%). The rock density of the samples was assumed to be 2.6 g/cm³ for all samples. All other required data (e.g., carrier data, mass dissolved, chlorine isotope ratios, etc.) to recalculate the natural ³⁶Cl concentrations and the age of the samples are given in the tables and Supplementary Table 1.

We sampled 11 boulders from moraine ridges on Mt. Barla (Tables 3.1 and 3.2; Figure 3.2a-c). The CRONUS Web Calculator version 2.0 (<http://www.cronuscalculators.nmt.edu>) (Marrero et al., 2016b) was used to calculate the sample ages. All ages have corrections for thickness and topographic shielding. We also applied a snow correction (~0.97) on spallation reactions with 50 cm snow thickness on top of sampled boulders assuming snow cover duration from December to April. Without snow correction, our ages would be >7% younger than those reported. We used the ³⁶Cl production rates reported in Marrero et al. (2021) [55.9 ± 4.5 atoms ³⁶Cl (g Ca)⁻¹ a⁻¹ for Ca spallation, 153 ± 12 atoms ³⁶Cl (g K)⁻¹ a⁻¹ K spallation and 696 ± 196 fast neutrons (g air)⁻¹ a⁻¹]. They were scaled following the time-dependent Lifton-Sato-Dunai method (also called “LSD” or “SF” scaling) (Lifton et al., 2014). Spallation and negative muon capture reactions by ⁴⁰Ca are responsible for >97% of total ³⁶Cl production, while lesser contributions are from ³⁹K (<0.4%) and slow neutron capture reactions by ³⁵Cl (~3%).

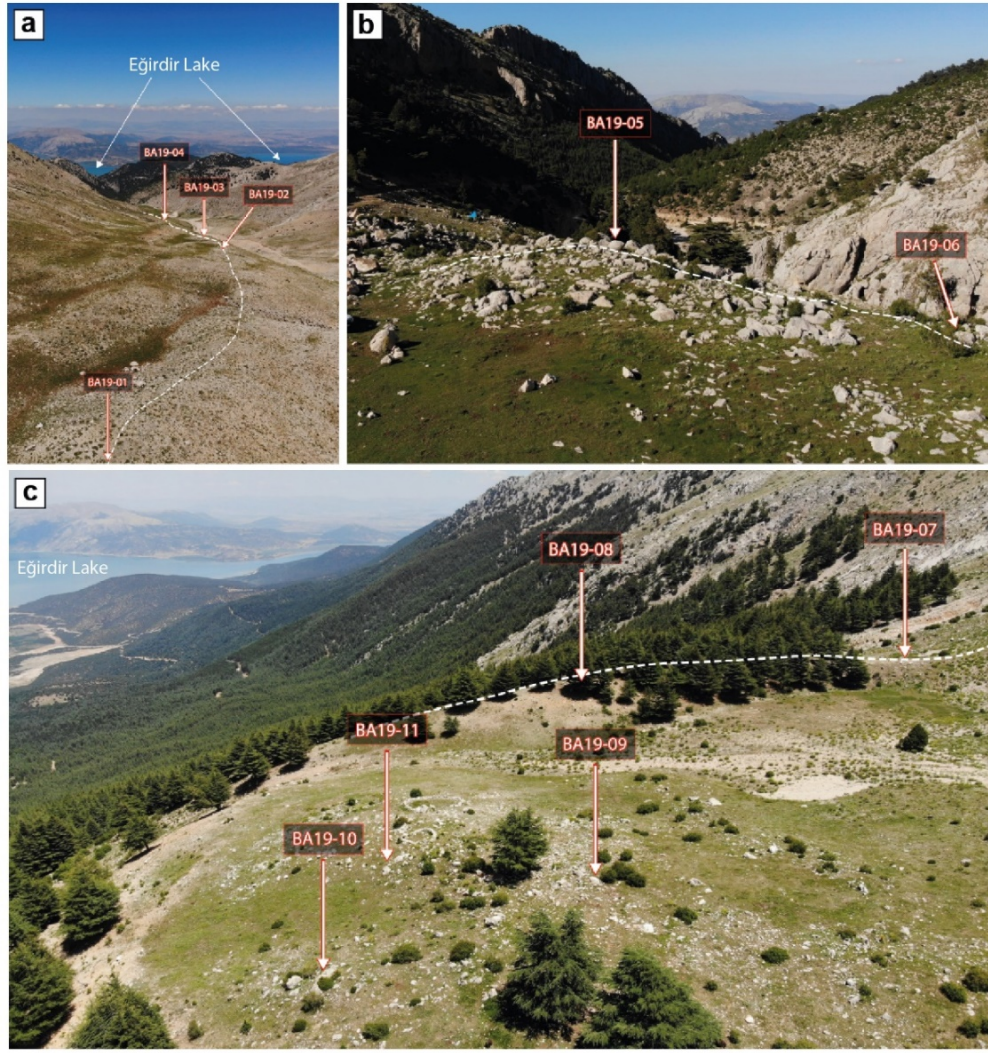


Figure 3.2 : Boulder locations on the moraine ridges: (a) the left lateral moraine and (b) the terminal moraine of Gedik Valley; (c) the outer terminal and right lateral moraines of Kapıdere Valley. Dashed-white lines indicate the moraine ridges. Note the absence of the ridgeline of the outer terminal moraine due to karstification and creep processes.

Table 3.2 : Geochemical data for the samples.

Sample ID	Major Elements											Trace Elements				
	Al ₂ O ₃	CaO	Fe ₂ O ₃	K ₂ O	MgO	MnO	Na ₂ O	P ₂ O ₅	SiO ₂	TiO ₂	CO ₂ (LOI)	Sm	Gd	U	Th	Cl
	(wt%)	(wt%)	(wt%)	(wt%)	(wt%)	(wt%)	(wt%)	(wt%)	(wt%)	(wt%)	(wt%)	(ppm)	(ppm)	(ppm)	(ppm)	(ppm)
BA19-01	0.15	55.07	0.06	0.03	0.27	0.05	<0.01	0.03	0.71	<0.01	43.5	0.49	0.74	<0.1	<0.2	3.9 ± 0.3
BA19-02	0.06	55.96	0.04	0.01	0.24	<0.01	<0.01	0.01	0.14	<0.01	43.5	0.13	0.21	1	<0.2	2.3 ± 0.2
BA19-03	<0.01	55.92	0.07	<0.01	0.4	<0.01	<0.01	<0.01	0.08	<0.01	43.5	<0.05	<0.05	1.4	<0.2	8.2 ± 0.5
BA19-04	0.28	54.44	0.17	0.06	0.4	0.02	0.02	0.05	1.19	0.01	43.2	0.54	1.05	<0.1	<0.2	9.3 ± 0.6
BA19-05	<0.01	55.35	0.18	<0.01	0.38	<0.01	<0.01	0.01	0.42	<0.01	43.5	<0.05	<0.05	1.1	<0.2	7.6 ± 0.5
BA19-06	<0.01	55.44	0.1	<0.01	0.37	<0.01	<0.01	<0.01	0.18	<0.01	43.8	<0.05	<0.05	1.1	<0.2	12.2 ± 0.7
BA19-07	<0.01	55.52	0.07	<0.01	0.53	<0.01	<0.01	0.01	0.15	<0.01	43.6	<0.05	<0.05	1.3	<0.2	7.1 ± 0.4
BA19-08	0.03	55.41	0.09	<0.01	0.62	<0.01	0.01	0.01	0.17	<0.01	43.6	<0.05	<0.05	1.2	<0.2	13.3 ± 0.8
BA19-09	0.02	55.65	0.1	<0.01	0.41	<0.01	<0.01	<0.01	0.15	<0.01	43.6	<0.05	<0.05	0.8	<0.2	10.9 ± 1.0
BA19-10	0.24	54.3	0.36	0.07	0.52	0.02	<0.01	0.05	0.98	<0.01	43.4	0.67	0.9	1.2	0.2	9.4 ± 0.5
BA19-11	0.05	55.37	0.12	0.01	0.53	<0.01	<0.01	<0.01	0.25	<0.01	43.6	<0.05	0.07	0.8	<0.2	14.8 ± 0.9

We calculated the ages with and without erosion correction. 10 mm ka⁻¹ of bedrock weathering was assumed for the erosion-corrected ages. In the Taurus Mountain Range, carbonate rock denudation rates were given in the range of ~5-15 mm ka⁻¹ (Hashemi et al., 2021). Thus, we preferred to use the erosion-corrected ages. All ages were reported in 1-sigma uncertainty, including the analytical and production rate errors (Figure 3.3a-c).

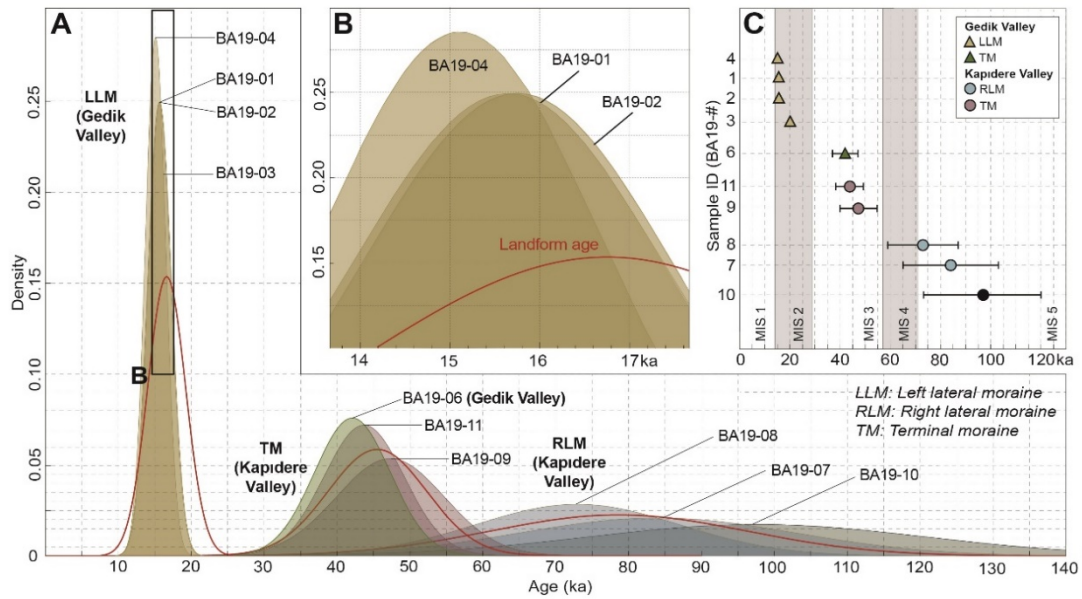


Figure 3.3 : Diagrams illustrating the cosmogenic ³⁶Cl surface exposure age distributions. (a) Color-filled camel plots represent the age of individual samples (n = 11). The best-estimate mean ages are shown in red lines for each landform calculated by Griffin et al. (2022). The terminal moraine age of Kapidere Gedik Valley (sample #6: 42.1 ± 5.2 ka) was given as a separate plot; (b) a closer view of the samples 1, 2, and 4 of the LLM of Gedik Valley; and (c) the distribution of all ages from Kapidere and Gedik valleys.

3.4 Results

3.4.1 Glacial geomorphology

3.4.1.1 Gedik valley

Gedik Valley is the largest valley of Mt. Barla, extending ~2.5 km from SW to NE on the eastern side formed by the convergence of two adjacent cirques at 2495 m a.s.l. (Figure 3.4). These cirques also constitute the head of their glacial valleys, namely Soğukkaynak and Ayıyalağı (Figures 3.1c and 3.4). Gedik Valley contains a large ground moraine at the intersection of these two valleys. A small rock glacier (Gedik Rock Glacier, 2300 m a.s.l., Figures 3.4 and 3.5) is present at this intersection point. The valley continues with two lateral moraines adjacent to the large ground moraine. The left lateral moraine extends parallel to the NW slope of Gedik Valley with an asymmetrical slope and is cut by several gullies. Its ridge appears at 2230 m a.s.l. and extends discontinuously for ~1.7 km down before terminating at 1950 m a.s.l. On the contrary, the right lateral moraine is not as well preserved as its left counterpart. Due to the deformed middle part of the right lateral moraine, it is difficult to follow the entire moraine ridge in the field (Figure 3.4). In front of these lateral moraines, a ground moraine appears intertwined with braided debris drainage at 1900 m a.s.l. There is also a tiny recessional moraine near the terminal moraine, which occupies a relatively small area compared with the other glacial landforms in the valley and does not display a clear ridge. We could not observe any glacial features below this point (~1850 m a.s.l.).

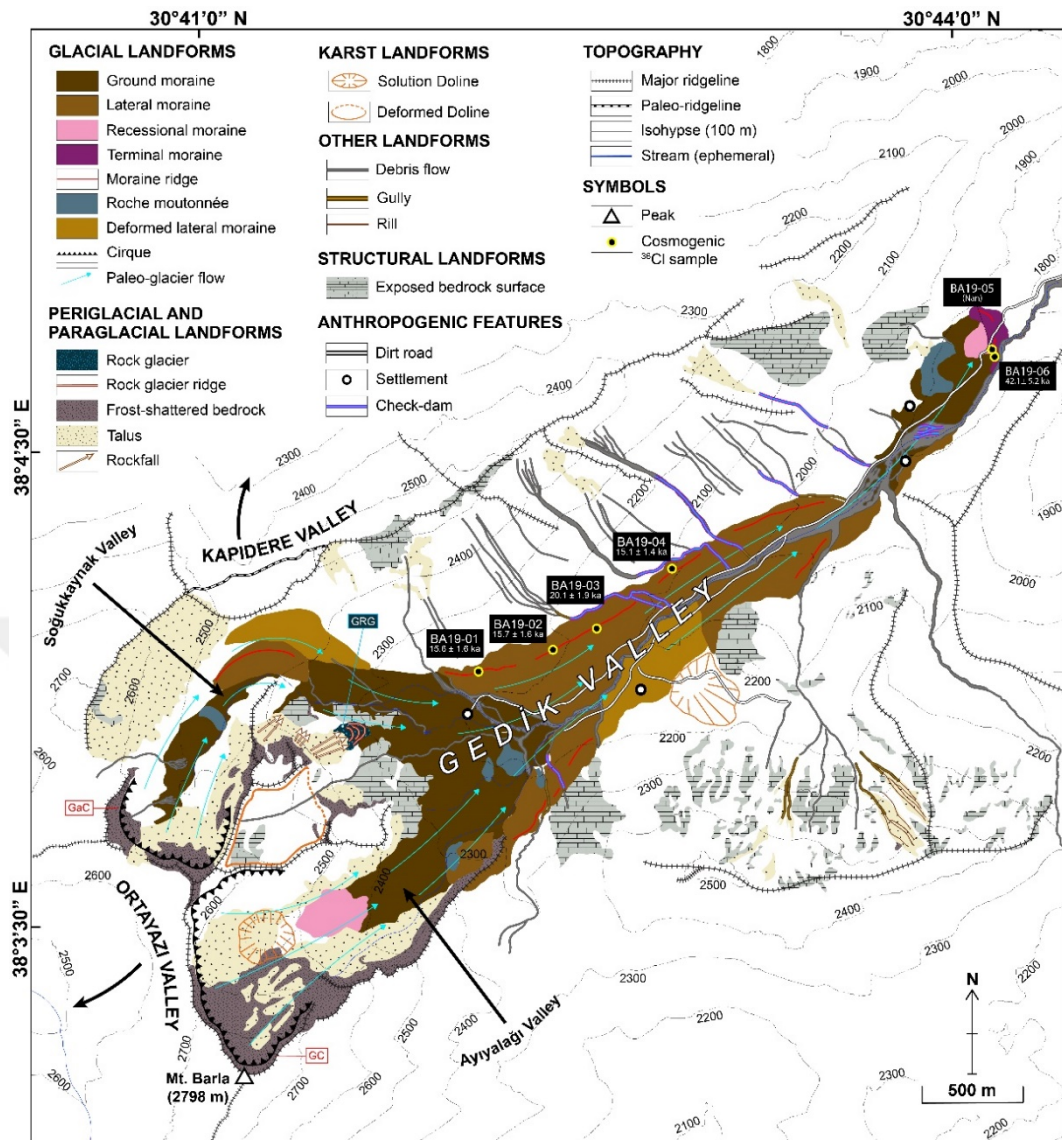


Figure 3.4 : Geomorphology map of Gedik Valley (GC: Gelincik Cirque; GaC: Gelincikana Cirque; GRG: Gedik Rock Glacier).

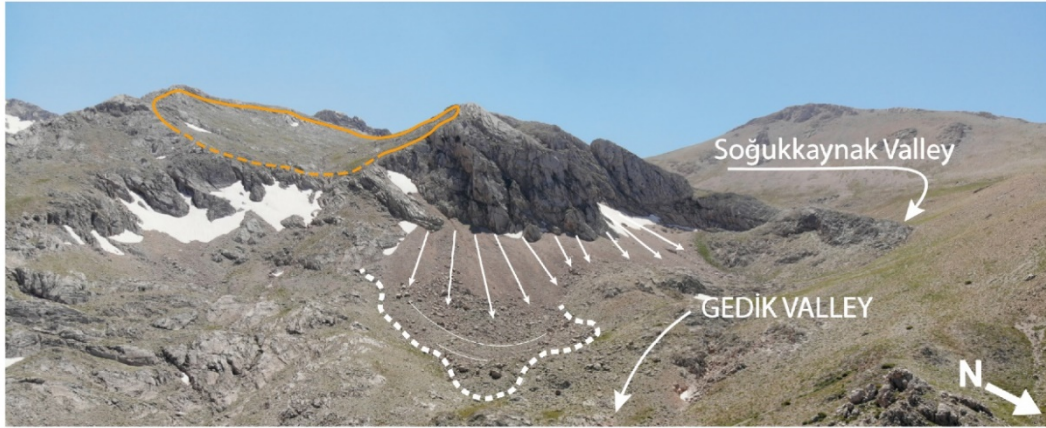


Figure 3.5 : Rock glacier located at the head of Gedik Valley. Thin white lines show the rock glacier ridges and white arrows on talus deposits represent rock falls. Dashed white lines indicate the margin of the rock glacier. The deformed doline between Gelincik and Gelincikana cirques is shown with the orange line. The dashed part of the line indicates the removed slope of the doline.

3.4.1.2 Kapidere valley

Kapidere Valley (Figures 3.1c and 3.6) is the only area studied by Ardos (1977a) on Mt. Barla. The valley is ~1.8 km long and oriented in an S-N direction with a W-E asymmetrical profile (Figure 3.7). The valley's eastern side exhibits gentler slopes than the western side.

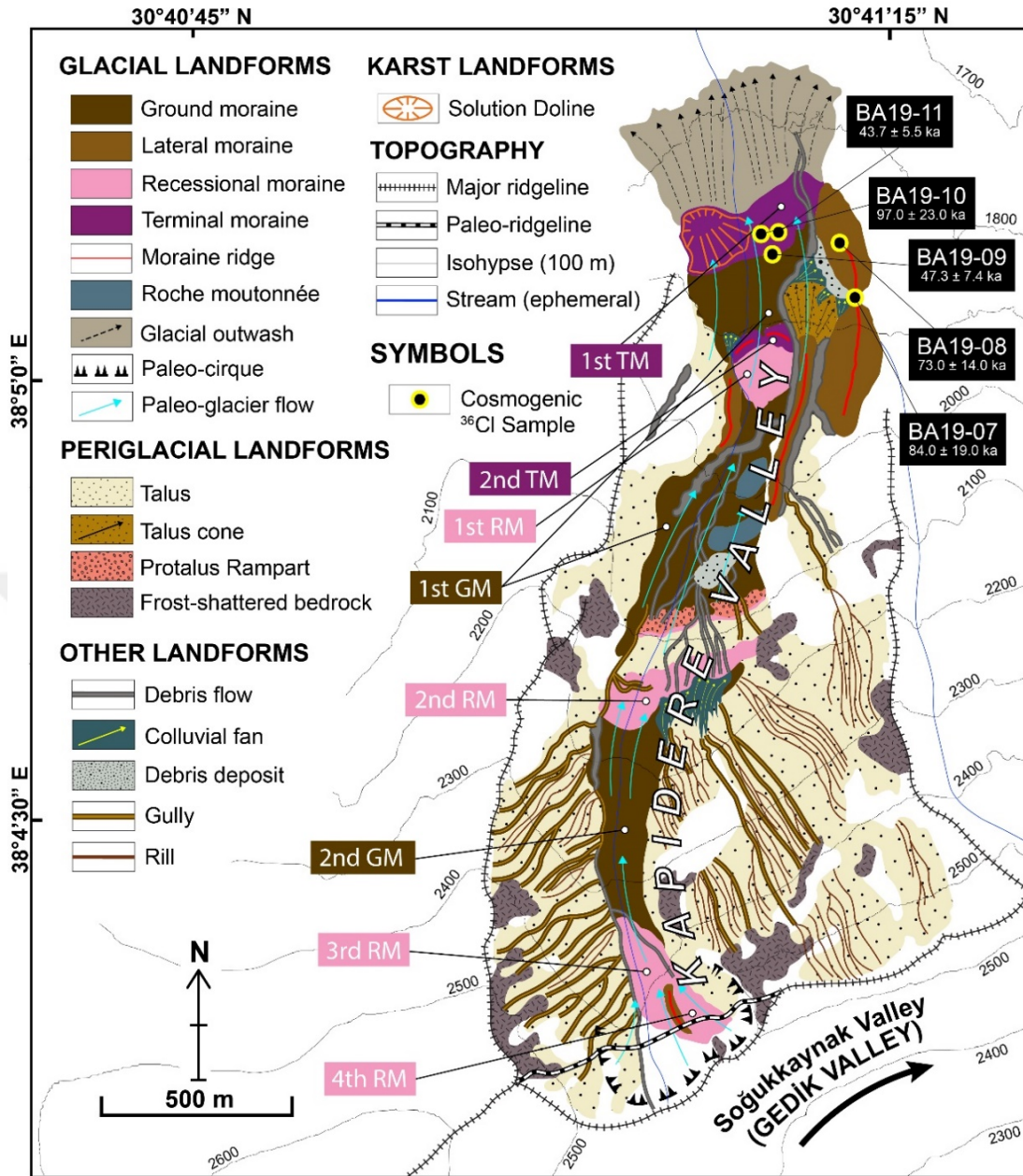


Figure 3.6 : Geomorphology map of Kapıdere Valley. Moraine formation orders from oldest to youngest are shown with sequence numbers (GM: ground moraine; RM: recessional moraine; and TM: terminal moraine).

There are several indications of Late Pleistocene glaciations in Kapıdere Valley (Figure 3.7). Two ground moraines were deposited on top of each other in the upper part of the valley. The first ground moraine has been exposed from the valley's center towards the terminal moraine (Figs. 3.6 and 3.7a). According to Ardos (1977a), the underlying first ground moraine has unsorted and angular Mesozoic limestone boulders, some of which are partly striated. The overlying second-ground moraine extends in the upper and western part of the valley as a tongue-shaped body. It is also

composed of Mesozoic limestones, but the sediments are not consolidated (Ardos, 1977a).

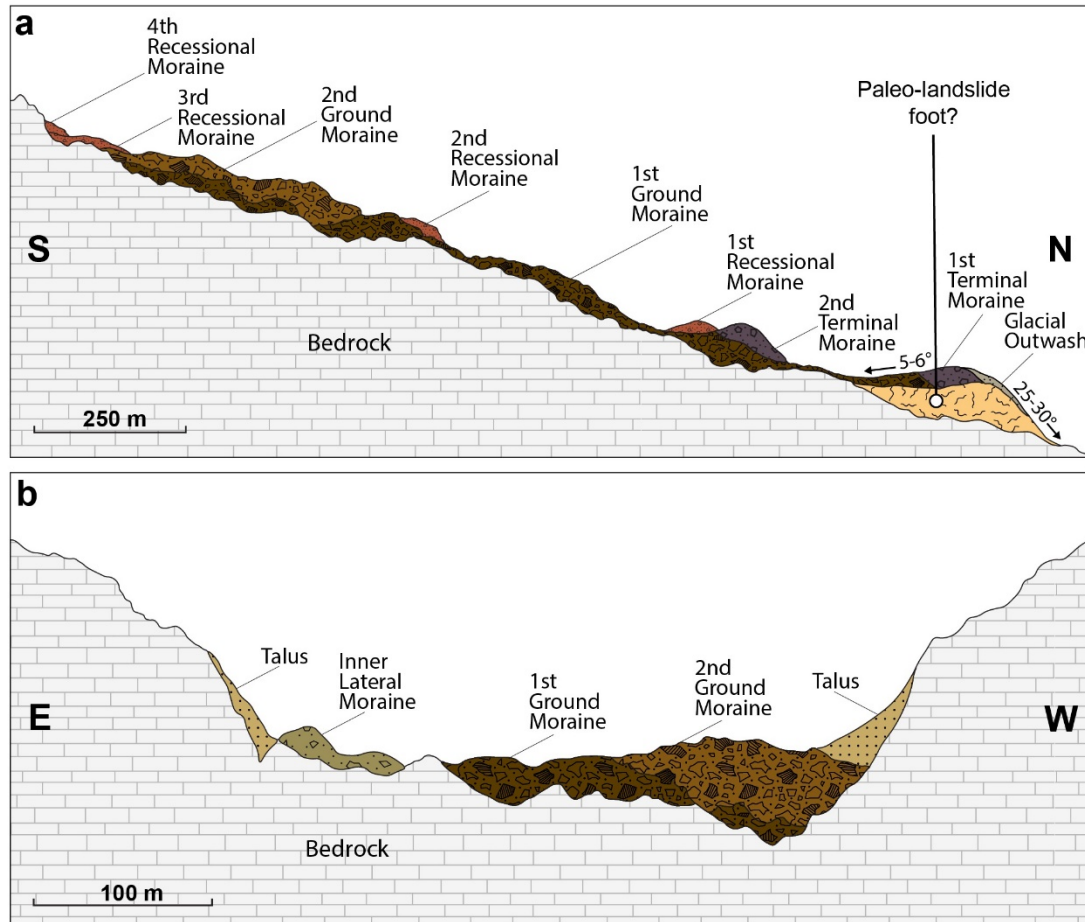


Figure 3.7 : (a) Longitudinal and (b) transverse sections of Kapıdere Valley (modified from Ardos, 1977a). Note the foot of a possible paleo-landslide underlying the moraines and glacial outwash.

Lateral moraines are found at different elevations as right and left lateral moraines. There are two right lateral moraines, the inner one at 1950 m a.s.l. and the outer one at 1835 m a.s.l., running along the end of the valley. The left lateral moraine (1950 m a.s.l.), which developed opposite to the inner right lateral moraine, has a shorter elongation (170 m) than the right lateral moraines (310 m for the inner right lateral moraine; 300 m for the outer right lateral moraine). A smaller left lateral moraine (2240 m a.s.l.) ~100 m long is located at the head of the valley.

The terminal moraines are the most deformed of this area's glacial deposits. The inner terminal moraine (1905 m a.s.l.) is located directly in front of the inner left and right lateral moraines. The moraine ridges are relatively distinguishable in the field. In contrast, the outer terminal moraine (1875 m a.s.l.) is more prominent and strongly

deformed and does not show a typical arcuate profile making it challenging to determine its ridge. This terminal moraine has a counter-dip of $5-6^\circ$ against the flow direction of the paleoglaciers and a dip of $20-25^\circ$ in the opposite direction (Ardos, 1977a) (Figure 3.7a). At the bottom of the valley where dense pine tree vegetation starts, the terminal moraine gradually turns into a glacial outwash. The bent trunks of the pine trees indicate a recent creep process on the slope of moraine and glacial outwash (Figure 3.8a, b).



Figure 3.8 : (a) Bent stems of the pine trees affected by creep processes on the glacial outwash overlying paleo-landslide foot; and (b) a rapid slope failure boulder on the glacial outwash of Kapıdere Valley under the dense pine vegetation.

Recessional moraines show a wide range of distribution throughout the valley (Figs. 3.6 and 3.7a, b). The inner terminal moraine bounds the first recessional moraine to the north and the inner lateral moraines from west to east. The second recessional moraine begins at the terminus of the second ground moraine. Colluvial fans, originating from talus deposits, overlie the moraines from SE. The third and fourth recessional moraines are located at the head of Kapıdere Valley. The third recessional moraine shows a longitudinal extension and narrows towards the western slope. The fourth recessional moraine is the smallest and developed on the third recessional moraine on the eastern slopes of the paleo-cirque (Figs. 3.6 and 3.7a).

3.4.1.3 Ortayazı valley

Ortayazı Valley shows slightly different glacial geomorphology than the other valleys (Figure 3.9). During the Late Pleistocene, glaciers formed three cirques developed on the SW slopes of the valley (e.g., Toklu, 2340 m a.s.l.; Akdere, ~2140 m a.s.l.; and İkizler, ~2450 m a.s.l.). Due to thick talus accumulations, we could not determine the

accurate elevations of the cirque floors of the Akdere and İkizler cirques. We are also considering the existence of a paleo-cirque at the head of the valley (Figure 3.9).

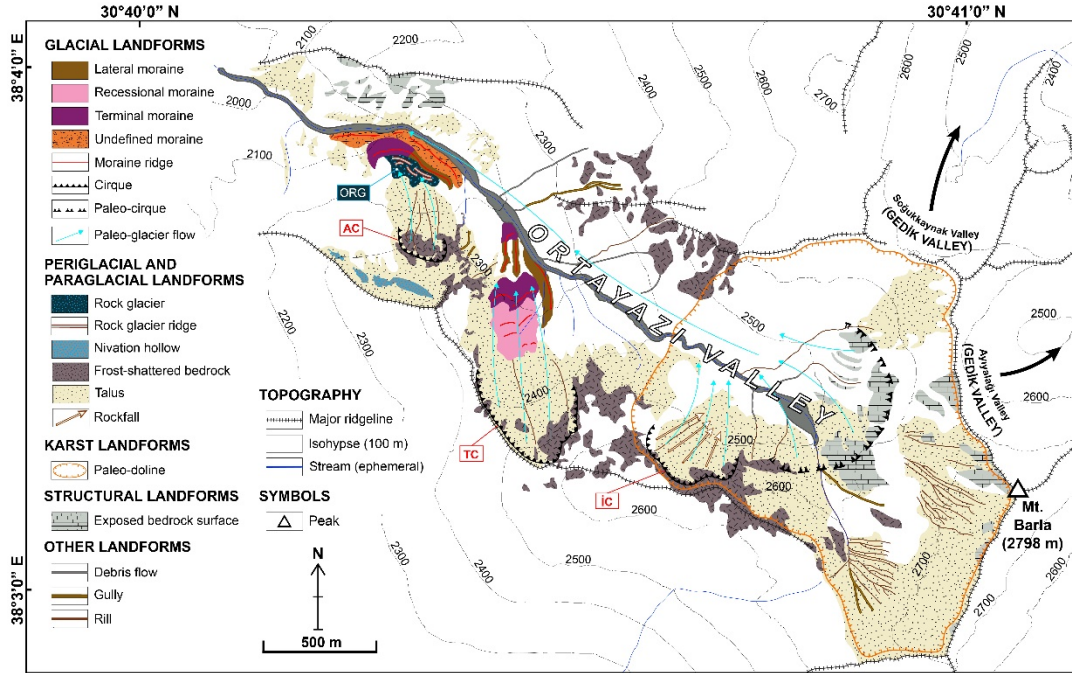


Figure 3.9 : Geomorphology map of Ortayazı Valley (İC: İkizler Cirque; TC: Toklu Cirque; AC: Akdere Cirque; ORG: Ortayazı Rock Glacier).

Among the cirques of Ortayazı Valley, Toklu Cirque (N-oriented) has a longitudinal circularity contrary to the other cirques in the valley. The Akdere Cirque (N-NNE-oriented) has the lowest altitude (~2140 m a.s. l.) among the other cirques on Mt. Barla. It has a rock glacier (Ortayazı Rock Glacier, 2125 m a.s.l., Figure 3.9) intertwined with talus deposits behind its terminal moraine. This rock glacier has patchy vegetation on its surface. Finally, İkizler Cirque (NE-oriented) is located proximally to the head of Ortayazı Valley. There are deformed moraine-like landforms in front of the cirque, but it is difficult to identify them contrary to other moraines that were distinguished in front of other cirques of the valley.

3.4.2 Glacial chronology

3.4.2.1 Gedik valley

We collected six samples from Gedik Valley (Figure 3.10a-f). Four samples (samples #1, #2, #3, and #4) come from the ridge of the left lateral moraine, the largest and most developed moraine in the valley (Figure 3.2a). From southwest to northeast, our samples yielded 15.6 ± 1.6 ka, 15.7 ± 1.6 ka, 20.1 ± 1.9 ka, and 15.1 ± 1.4 ka,

respectively. Two different glaciations could be inferred based on these ages. First, using the weighted average (15.5 ± 0.9 ka) of three closer results (samples #1, #2, and #4), a Late-glacial landform age could be tentatively proposed. However, the apparent age of the moraine might be younger than its true age since the moraine ridge settles over time, causing exhumation and/or boulder toppling (Hallet & Putkonen, 1994; Zreda et al., 1994; Ivy-Ochs & Briner, 2014). In Gedik Valley, the position of the left lateral moraine is adjacent to the NW slope of the valley. Hence, some of the sampled boulders might have been affected by slope processes and/or exhumed due to postglacial processes. Contrary to this, boulders #1 and #3 are well-rooted into the moraine matrix. Putkonen & Swanson (2003) claimed that the oldest boulder age would be 90% of the true age of the moraine with a 95% probability. However, when considering these perfectly overlapping three ages above (15.6 ± 1.6 ka, 15.7 ± 1.6 ka, and 15.1 ± 1.4 ka), using the ‘oldest age’ (20.1 ± 1.9 ka) for landform age does not seem reasonable. Hereby, we preferred to use an alternative method described by Griffin et al. (2022, in review) to represent the age of the left lateral moraine. Based on this method, we combined all ages (15.6 ± 1.6 ka, 15.7 ± 1.6 ka, 20.1 ± 1.9 ka, and 15.1 ± 1.4 ka) to find the most probable age. In this way, owing to the more significant uncertainty, we found a more coherent landform age for the left lateral moraine of Gedik Valley. Accordingly, the mean and standard deviation of the combined samples from all these ages yielded the landform age as 16.6 ± 2.6 ka (Figure 3.3a-c).



Figure 3.10 : Pictures of the sampled boulders for the cosmogenic surface exposure dating: (a–d) left lateral moraine of Gedik Valley; (e) and (f) terminal moraine of Gedik Valley; (h) and (i) outer right lateral moraine of Kapidere Valley; (j–k) outer terminal moraine of Kapidere Valley. White arrows indicate the sampling points on the boulders.

In addition, we collected two more samples (#5 and #6) from the terminal moraine ridge (Figure 3.2b). Unfortunately, the ages obtained from the two samples were

insufficient to determine the landform age. Moreover, we could not calculate the erosion-corrected age of sample #5 due to the saturation of ^{36}Cl and considered an outlier. Although it is not adequate to date the terminal moraine based on a single age (sample #6: 42.1 ± 5.2 ka), it is interesting to note that it yields a comparable chronology to Kapıdere Valley, given below. Consequently, we reported the age of sample #6 as the terminal moraine age, which indicates the oldest (pre-LGM; MIS 3) glacial retreat in the valley (Table 3.3).

Table 3.3 : Measured ^{36}Cl inventories and exposure ages of boulders and weighted average ages of landforms. TCND outliers are indicated with an asterisk (*) at the end of the sample IDs.

		³⁶ Cl age			
Sample ID	Landform	³⁶ Cl (measured)			Landform age
		(10 ⁴ atoms g ⁻¹ rock)	without erosion correction (ka)	erosion corrected (10 mm ka ⁻¹) (ka)	
Gedik Valley					
BA19-01	Left Lateral Moraine	130.1 ± 8.0	14.0 ± 1.3	15.6 ± 1.6	16.6 ± 2.6
BA19-02	Left Lateral Moraine	129.4 ± 8.0	14.0 ± 1.3	15.7 ± 1.6	
BA19-03	Left Lateral Moraine	165.0 ± 9.5	17.6 ± 1.6	20.1 ± 1.9	
BA19-04	Left Lateral Moraine	117.8 ± 6.8	13.7 ± 1.2	15.1 ± 1.4	
BA19-05*	Terminal Moraine	614.9 ± 37.0	79.3 ± 9.1	NaN ± NaN	42.1 ± 5.2
BA19-06	Terminal Moraine	244.0 ± 13.6	30.8 ± 2.7	42.1 ± 5.2	
Kapıdere Valley					
BA19-07	Right Lateral Moraine	365.2 ± 20.6	44.2 ± 4.1	84.0 ± 19.0	78.5 ± 17.6
BA19-08	Right Lateral Moraine	359.7 ± 19.8	42.7 ± 3.8	73.0 ± 14.0	
BA19-09	Terminal Moraine	279.4 ± 16.4	34.0 ± 3.0	47.3 ± 7.4	45.5 ± 6.8
BA19-10*	Terminal Moraine	384.8 ± 21.8	47.0 ± 5.0	97.0 ± 23.0	
BA19-11	Terminal Moraine	267.9 ± 15.2	32.8 ± 2.9	43.7 ± 5.5	

3.4.2.2 Kapıdere valley

We collected five samples from Kapıdere Valley (#7, #8, #9, #10, and #11) (Figure 3.2c). Two of them (samples #7: 84.0 ± 19.0 ka and #8: 73.0 ± 14.0 ka) were sampled from the outer right lateral moraine ridge (Figure 3.10g, h). The remaining three samples (#9: 47.3 ± 7.4 ka, #10: 97.0 ± 23.0 ka, and #11: 43.7 ± 5.5 ka) were collected from the ridge of the outer terminal moraine (Figure 3.10i-j).

The inconsistent results make it difficult to evaluate the glacial chronology of the Kapıdere Valley based on the ^{36}Cl surface exposure ages. We presume that the postglacial processes have affected the outer terminal moraine. For instance, the karstic depressions (e.g., solution dolines; Figure 3.6) within the boundaries of the

external terminal moraine might have moved the boulders since the karstification started. It is known that karstification might modify the original positions of glacial deposits (Alonso, 1998). Besides, its topography might have been reworked by creep processes (Figure 3.8a). However, the obtained ages from the terminal moraine of Kapıdere Valley (samples #9: 47.3 ± 7.4 ka and #11: 43.7 ± 5.5 ka) are consistent with the terminal moraine of Gedik Valley (42.1 ± 5.2 ka), except for sample #10 (97.0 ± 23.0 ka) considered an outlier probably because of inherited ^{36}Cl from previous exposure to cosmic radiation (Table 3.3). Based on the mean and standard deviation of two meaningful ages (samples #9 and #11), we found that the outer terminal moraine age is 45.5 ± 6.8 ka (Table 3.3). Considering the terminal moraines from the Gedik and Kapıdere valleys, we conclude a pre-LGM (MIS 3) glaciation.

The outer right lateral moraine ages (samples #7: 84.0 ± 19.0 ka and #8: 73.0 ± 14.0 ka) provide the oldest results in the study area. The mean and standard deviation of these two samples yields 78.5 ± 17.6 ka, indicating an earlier glacial advance during MIS 5 (Table 3.3). This age is broadly consistent with sample #10 (97.0 ± 23.0 ka), which we previously considered an outlier but much older than the other samples in the whole of Mt. Barla. On the other hand, although sample #10 is located within the boundaries of the terminal moraine, it may have been detached from the right lateral moraine due to the paleoglacier advance/retreat in the valley throughout the ~45 ka event. Then, it finally settled within the outer terminal moraine after the glaciers disappeared. Thus, sample #10 may be interpreted as a part of the right lateral moraine along with samples #7 and #8. In this scenario, the mean and standard deviation of these three samples yields an age of 84.6 ± 21.3 ka for the right lateral moraine. Since we have no evidence to support this claim other than the older sample age, we preferred to use the 78.5 ± 17.6 ka as the formation age of the right lateral moraine.

3.5 Discussion

3.5.1 Evolution of glacial landforms

3.5.1.1 Gedik and Ortayazı valleys

On Mt. Barla, the cirques of Gedik Valley (Gelincikana and Gelincik cirques) started to develop during the fluvial to karst domination transition (Figure 3.11a-c). Karst depressions (e.g., dolines and kontas) are suitable morphological features for cirque

development (Fels, 1929; Žebre & Stepišnik, 2015; Öztürk et al., 2018, 2021). Pre-glacial karst depressions are reshaped by glaciation (Lewis, 1949, 1960; Veress et al., 2019) and are referred to as kontas (Žebre & Stepišnik, 2014, 2015). A deformed doline (~2500 m a.s.l.) (Figures 3.4 and 3.5) located between Gelincikana and Gelincik cirques suggests glaciokarst-based cirque development is possible on the summit of Mt. Barla. Gelincik Cirque also has a solution doline that deepens its floor (Figure 3.4). This relationship is defined as karstiglacial and karstic adaptations of glacial closed depressions in bedrock (Ford, 1979). The dip of the strata on the slopes could also provide a suitable environment for karst development, as the strata dipping in opposite directions to the surface drainage accelerates karst formation (Bayrakdar, 2012).

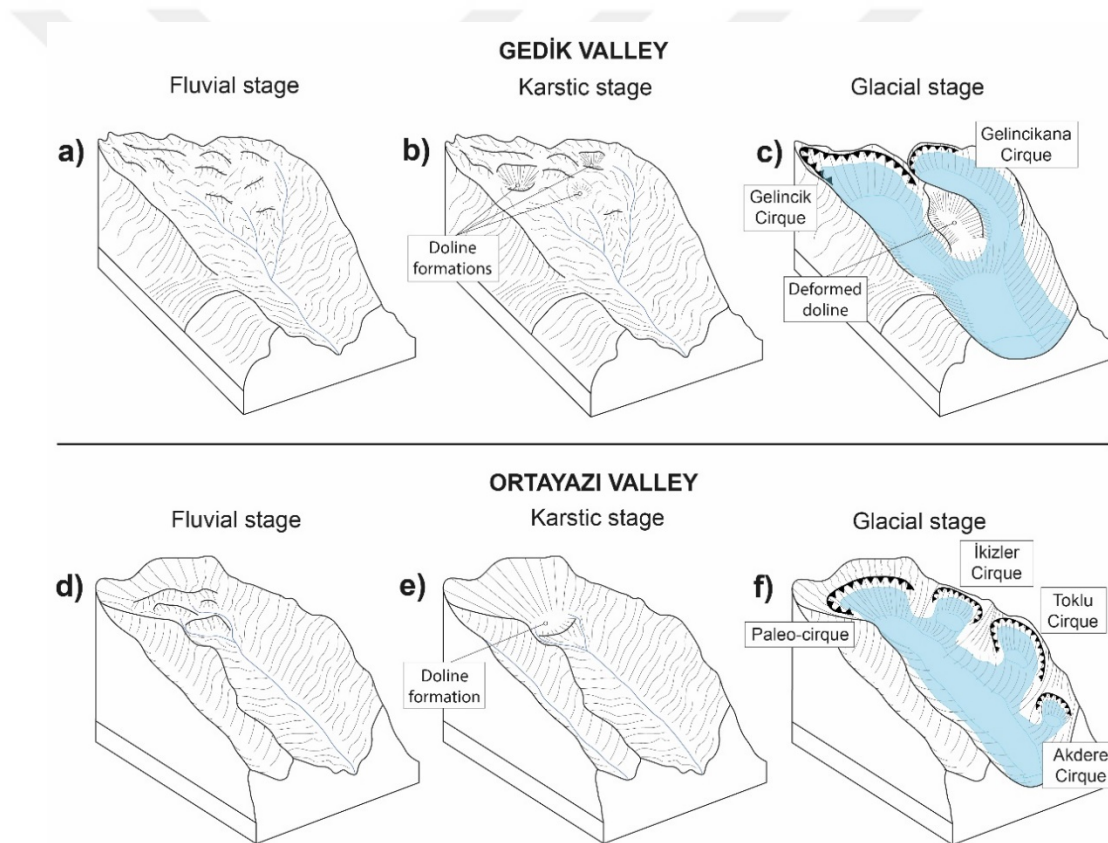


Figure 3.11 : Sketches of the dominant stages in the evolution of Gedik (a–c) and Ortayazı (d–f) valleys: (a, d) fluvial and karstic processes are active; (b, e) karstic processes are dominating: doline development near hillslopes; (c, f) glaciokarstic cirque development in the glacial stage.

The head of the Ortayazı Valley, adjacent to Gelincikana Cirque from the northeast and Gelincik Cirque from the east, formed similar to cirques in Gedik Valley within the context of glaciokarstic evolution (Figure 3.11d-f). The glaciation of the Ortayazı

Valley might have started within the main cirque (paleo-cirque, ~2550 m a.s.l.), much earlier than the glaciations on the smaller cirques on the southwestern slope of the valley. The terminal moraines of the Akdere and Toklu cirques corroborate that valley paleoglaciers developed before cirque paleoglaciers. Following the valley paleoglacier retreat, cirque paleoglaciers were formed. Otherwise, valley paleoglacier would have removed the terminal moraines as a younger glacial advance. Although we could not determine a preserved cirque on the head of the valley, we infer that there should have been a paleo-cirque based on a deformed cirque headwall-like landform. We assume that this cirque might have been developed in a massive paleo-doline or konta (Žebre & Stepišnik, 2015) with a diameter of ~1.5 km. The İkizler Cirque, at the beginning of the valley, is the closest cirque to the paleo-cirque. However, we could not find clear evidence about moraine development in front of this cirque. İkizler Cirque might have been developed in the same period as the paleo-cirque, and later, its moraines were removed by erosion.

3.5.1.2 Kapidere valley

Kapidere Valley represents a distinct and more complex morphological development than other glaciated valleys (Figure 3.12a-d). The position, age, and morphology of moraines evidence this peculiar geomorphology. We propose two scenarios regarding its origin: (1) It might have developed as a subsidiary valley from Soğukkaynak Valley, or (2) it formed from its paleo-cirque (Figs. 3.12c and 3.13a).

KAPIDERE VALLEY

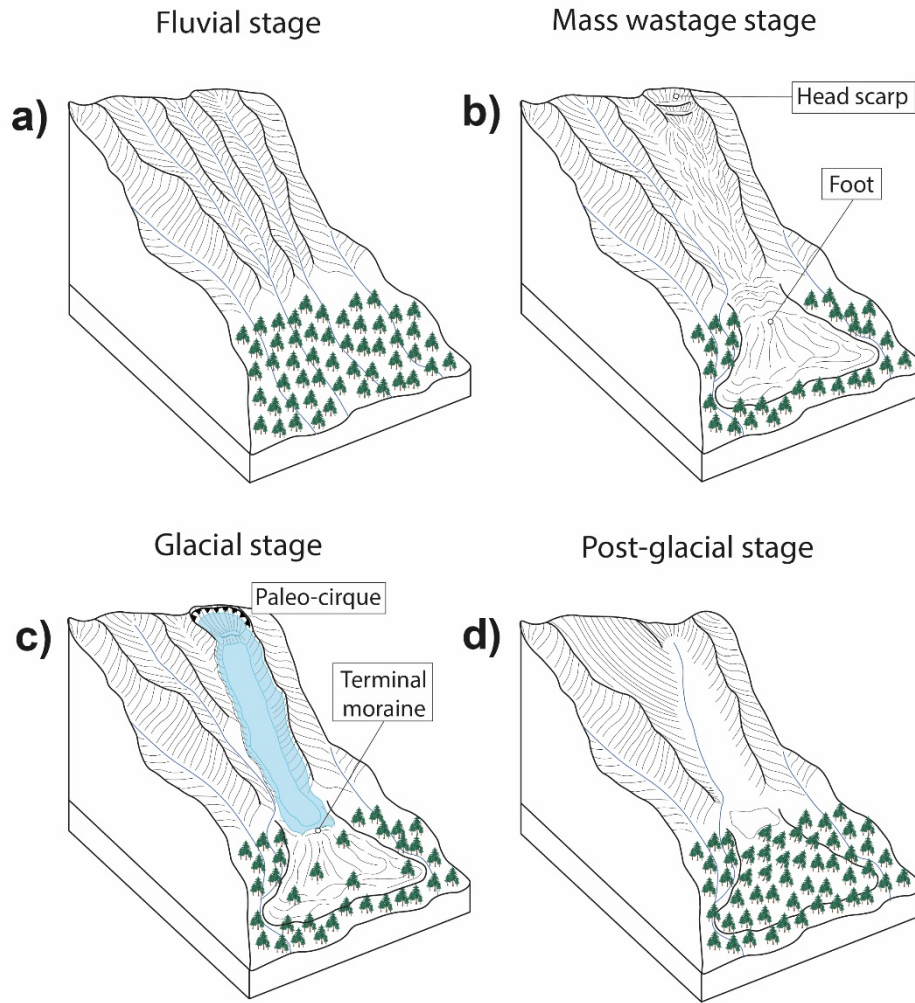


Figure 3.12 : Sketch of the dominant stages in the evolution of Kapıdere Valley: (a) fluvial processes are dominant; (b) paleo-landslide; (c) cirque development in the head scarp of the landslide during the glacial stage; (d) recent topography after the cirque headwall failure (i.e., rapid slope failure).

Kapıdere Valley appears as a subsidiary valley separated from Soğukkaynak Valley at first glance. However, the convexity of the two valleys shows that Kapıdere Valley could not have been developed by the Soğukkaynak paleoglacier (Figs. 3.1c and 3.13a). In other words, it must have had its proper cirque. To clarify this assumption, we propose two observations:

(i) Soğukkaynak Valley starts to extend to the north and turns towards the east (Figs. 3.1c and 3.13a) at the head of Kapıdere Valley. On the other hand, the Kapıdere Valley has an N-NNW development and then turns to N-NNE.

(ii) If a paleoglacier advance formed Kapıdere Valley throughout Soğukkaynak Valley, there should not have been the threshold at the head of Kapıdere Valley. In other words, this threshold should have been removed by Soğukkaynak paleoglacier.

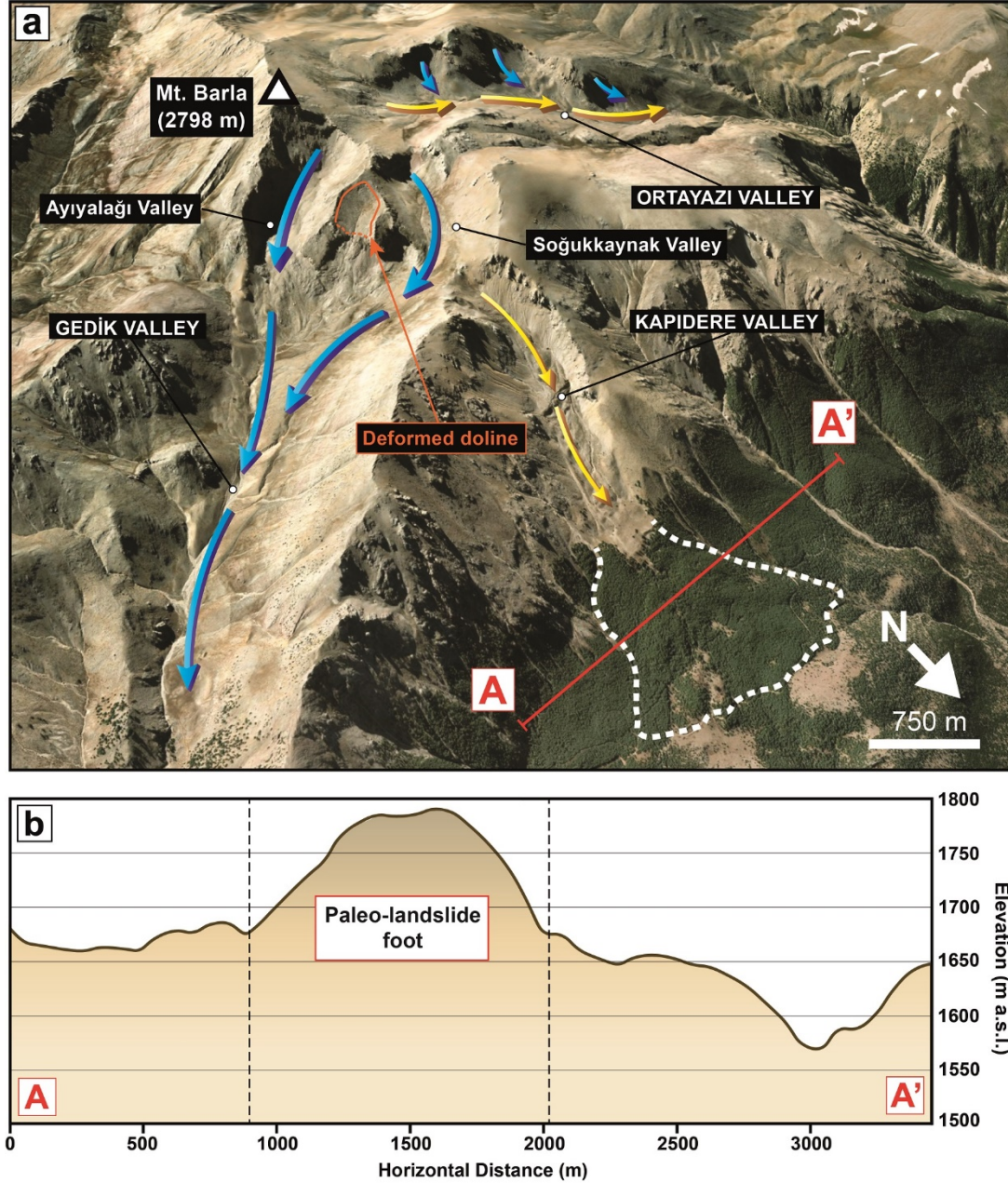


Figure 3.13 : (a) Paleoglacier flow directions on the valleys of Mt. Barla and transversal section line of the paleo-landslide foot (dashed white line) at the end of the Kapıdere Valley. Arrows indicate the cirque-based (blue) and paleo-cirque-based (yellow) paleoglacier flows; (b) transversal section graph of the paleo-landslide foot. Dashed lines show the foot boundaries.

Therefore, the direction and convexity of Kapıdere and Soğukkaynak valleys and their boundaries indicate a paleo-cirque development at the head of Kapıdere Valley (Figs. 3.1c and 3.13a). The cirque headwall was presumably removed by a rapid slope failure

(i.e., paraglacial response) and then transformed into its recent form. The large boulders on the glacial outwash surface covered by dense pine vegetation (Figs. 3.12d and 3.13a) might indicate a rapid slope failure based on paraglacial response (Figure 3.8a-b). A comparable example of an instantaneous slope failure-cirque relationship was studied by Jarman & Lukas (2007), who described a sizeable postglacial rockslide that affected a cirque in Ben Hee in the Northern Highlands (Scotland) and caused part of the headwall to collapse. Another example of a cirque headwall-originated rock avalanche has been recently studied in Mt. Geyikdağ (Central Taurus Range), where the avalanche has traveled ~1.8 km (Hashemi et al., 2022).

Besides a cirque headwall-based rapid slope failure, a paleo-landslide might have been part of the polygenetic evolution of Kapıdere Valley. According to Tunusluoglu et al. (2008), mass movements like debris flows are frequent on the northern slopes of Mt. Barla. As cirques could develop from landslide scars (Derbyshire & Evans, 1976) the development of the valleys might have started with a landslide on a non-glacial topography during its fluvial stage (Figure 3.12a, b). Following the emplacement of this hypothetical landslide, suitable conditions were provided by the main scarp for formation of the glacier. Later, the cirque (i.e., paleo-cirque) of Kapıdere Valley was developed, and the valley glacier grew (Figure 3.12c). The cirque headwall collapsed once the glacier disappeared because of a paraglacial response. Eventually, the valley attained its current morphology since the postglacial processes started (Figure 3.12d).

Given the relative thickness of the accumulated material at the end of the valley, we consider that the foot of the hypothetical paleo-landslide might have been underlined by the outer terminal moraine (Figures 3.7a and 3.13b). We also observed a broad fan-shaped morphology that starts to extend from in front of the outer terminal moraine to the lower slopes of Mt. Barla throughout 1.1 km (Figure 3.13a). This fan-shaped morphology is most likely formed by reworked material on the foot of the hypothetical paleo-landslide. Otherwise, it seems impossible that the material accumulated due to a single glaciation event during its maximum extent at 78.5 ± 11.8 ka. Even though a glacial outwash (Figure 3.7a) covers the highest part of this fan-shaped deposit, we are unsure of its lower limits because of dense vegetation. The light detection and ranging (LiDAR) method may provide insight into a landscape covered by lush vegetation overlying the aforementioned fan-shaped deposit (Görüm, 2019).

Although we did not perform any analysis to determine whether there was a landslide or not before the glacial history of the valley, specific methods could be applied to check the hypothetical presence of a paleo-landslide in Kapıdere Valley. Methods such as slope variability and Terrain Ruggedness Index (TRI) analyses (Popit & Verbovšek, 2013) produce engineering geomorphology maps and measure trends of lineaments and orientations of mapped faults (Francioni et al., 2018). Borehole applications, resistivity surveys, and seismic surveys (Ersoy et al., 2019) might also show paleo-landslides and their characteristics.

3.5.2 Comparison with other glacial chronologies in the Taurus mountains

We provide cosmogenic ^{36}Cl surface exposure age data for at least two and possibly three deglaciations on Mt. Barla between MIS 5 and MIS 2. Except for the MIS 5, similar chronologies were reported from surrounding mountains in the Taurus Mountain Range (see Sarıkaya & Çiner, 2017 and references therein; Akçar, 2022b). For instance, the oldest pre-LGM (MIS 3) ages come from Mt. Akdağ (38.9 ± 0.8 ka; Sarıkaya et al., 2014) and Bolkar Mts. (46.0 ± 7.0 ka; Çiner & Sarıkaya, 2017). The LGM extent of most of the mountains in the Taurus Range is well established, and all mountains contain different types of moraines developed during MIS 2. From west to east these include Mt. Sandıras (19.6 ± 2.0 ka; Sarıkaya et al., 2008), Mt. Akdağ (20.3 ± 0.4 ka and 19.7 ± 0.5 ka; Sarıkaya et al., 2014), Mt. Dedegöl (29.8 ± 2.2 ka and 22.0 ± 1.7 ka; Zahno et al., 2009; 26.7 ± 0.9 ka and 18.4 ± 0.7 ka; Köse et al., 2019), Mt. Geyikdağ (19.5 ± 0.2 ka; Çiner et al., 2015; 19.1 ± 0.8 ka; Sarıkaya et al., 2017) and Bolkar Mts. (25.0 ± 4.1 ka and 18.9 ± 3.3 ka; Çiner & Sarıkaya, 2017).

Our knowledge about the pre-LGM glaciations of the Taurus Mountains (for the whole of Anatolia as well) is quite limited. In general, the oldest glaciations go back to MIS 3. On the other hand, the MIS 5 ages (84.0 ± 19.0 ka and 73.0 ± 14.0 ka) of Mt. Barla give another perspective to understanding the region's glacial history. Late-glacial landforms developed by glaciers cover larger areas than pre-LGM glacial landforms. Mostly, they might have removed or deformed the previous glacial evidence. The MIS 5 ages of Mt. Barla belong to the external right lateral moraine of Kapıdere Valley. This valley also has MIS 3 ages from the terminal moraine, adjacent to the right lateral moraine. Therefore, we can infer that the MIS 5 glaciations were more extensive than the MIS 3 glaciations in Kapıdere Valley.

Moreover, the inner terminal and lateral moraines on the higher part of the valley presumably belong to LGM or younger glaciations. Alternatively, we have pre-LGM and Late-glacial ages (42.1 ± 5.2 ka for the terminal moraine and 16.6 ± 2.6 ka for the left lateral moraine) in the Gedik Valley. It seems that the paleoglaciers of Mt. Barla reached their maximum extent at MIS 3. Correspondingly, the development of Late-glacial paleoglaciers has remained limited in contrast to pre-LGM paleoglaciers.

On the other hand, we did not encounter Holocene advances on Mt. Barla in contrast with the other sites along the Taurus Mountain Range (Sarıkaya & Çiner, 2015; Altınay et al., 2020; Oliva et al., 2020). It is possible that we did not sample younger moraines, presuming they existed. Another possibility might be related to the location of the mountain, situated in the northernmost part of the Western Taurus at a transition zone from the Mediterranean sub-humid to Mediterranean semi-arid climatic zones (Akman & Ketenöğlü, 1986). Presumably, glaciers could not find an opportunity to advance after the Late-glacial due to arid climate conditions. Although the crescent-shaped (Figs. 3.1c and 3.4) left lateral moraine on Soğukkaynak Valley might indicate a younger advance, quantitative data are needed to confirm this assumption. Today, there are no present glaciers on the mountain.

3.6 Conclusion

We provide a glacial geomorphology map of three valleys in Mt. Barla (SW Türkiye), where karstic depressions and mass movements influenced the cirque developments resulting in polygenetic and polycyclic glacial landforms. We also report multiple glaciations based on geomorphological evidence and numerical dating from lateral and terminal moraines. Cosmogenic ^{36}Cl surface exposure dating based on 11 moraine boulders indicates at least two and possibly three deglaciations during the Late Pleistocene. In the Gedik Valley, while the left lateral moraine, dated to 16.6 ± 2.6 ka, indicates the midst of Late-glacial (MIS 2), the terminal moraine yielded the most extensive glaciation at 42.1 ± 5.2 ka (MIS 3). The outer terminal moraine age (45.5 ± 6.8 ka) of the Kapıdere Valley confirms this pre-LGM (MIS 3). An earlier glaciation in the Kapıdere Valley at 78.5 ± 17.6 ka (MIS 5) is possible, even though more data are needed to confirm this extent.



4. A RARE PIEDMONT GLACIATION IN THE MEDITERRANEAN: INSIGHTS FROM COSMOGENIC ³⁶Cl DATING OF DAVRAZ HUMMOCKY MORAINES FIELD (SW TÜRKİYE)³

4.1 Introduction

In various regions worldwide, particularly since the global Last Glacial Maximum (LGM; 26.5 to 19 ka, Clark et al., 2009), the growth of specific glaciers has been limited by the topography of the mountainous areas in which they originated. However, some glaciers surpass these boundaries and extend onto relatively flat terrain, spreading out like a colluvial fan. This transition marks their shift to a distinct spatial existence, leading to their classification as piedmont glaciers (or lobes). Valley glaciers that flow through bedrock troughs and terminate in low-lying regions can give rise to these types of glaciers (Benn & Evans, 2010). They may either have a cirque glacier origin (e.g., Smith et al., 1997), or they may originate from ice caps or ice fields (e.g., Evans et al., 2016; Ivy-Ochs, 2015), like the famous Malaspina Glacier in Alaska, USA (Post, 1972; Sharp, 1958); the largest piedmont glacier on the earth, covering an area of 3209 km² (Kargel et al., 2014). Piedmont glaciers also exist on different planets than Earth, such as the Late Amazonian piedmont glacier located northwest of Olympus Mons on Mars (Milkovich et al., 2006). Several piedmont glaciers may merge and thicken to form small ice fields, which in turn merge to form larger ones. They thicken and grow due to their low gradients and ice velocities, which prevent them from discharging the ice from fast-moving valley glaciers that originate from steep mountainous areas (Bennett & Glasser, 2009). After spreading out onto a flat area, piedmont lobes form frontal moraines within preexisting end moraine systems that were established during earlier ice advances (Ivy-Ochs et al., 2022). They provide great examples of large hummocky moraine fields, such as the ones in the Northern Alpine Foreland (Salzach Glacier) (Salcher et al., 2010), on Geyikdağ,

³ This chapter is based on the paper “Altınay, O., Sarıkaya, M. A., & Wilcken, K. M. (2024) A rare piedmont glaciation in the Mediterranean: Insights from cosmogenic ³⁶Cl dating of Davraz hummocky moraine field (SW Türkiye). *Earth Surface Processes and Landforms*, 1-20.”

Taurus Mountains, Türkiye (Çiner et al., 2015) and on the Rhine and Reuss glacier systems (Kamleitner et al., 2023).

Hummocky moraines are primarily characterized as uncontrolled ice-disintegration features that are predominantly irregular and exhibit no apparent elongation (Gravenor & Kupsch, 1959). They are formed by a seemingly irregular collection of morainic mounds and ridges, giving the appearance of a chaotic or disorganized assemblage of glacial deposits (Balfour, 2014). This definition may include a diverse range of landforms, which may have various origins such as deposition linked to active or stagnant (i.e., dead) ice, and deformation due to glaciotectonic activity (Benn & Evans, 2010). Hambrey (1994) described hummocky moraine origin as a down-wastage of a glacier, rather than its recession, particularly during a state of glacial inactivity. Correspondingly, the term ‘hummocky moraine’ has been used to refer to glacial sediments deposited during the down-wasting of debris-mantled glaciers.

Piedmont glacier-originated hummocky moraines can be seen near large ice sheets and in valleys in high latitudes (e.g., Astakhov, 2018; Briner et al., 2005). Additionally, it is possible to encounter them at smaller scales in mountainous regions in mid-latitudes (e.g., Çiner et al., 2019; Karlstrom, 1999; Salcher et al., 2010). Certain dating studies have been conducted on depositional landforms formed by the piedmont glaciers (e.g., Blaise et al., 1990; Briner et al., 2005; Dong et al., 2022; Reuther et al., 2011; Zech et al., 2005a, 2005b). Besides terminal, latero-terminal, and end moraines, hummocky moraines are also suitable for dating the retreat of (paleo-)piedmont glaciers (e.g., Salcher et al., 2010; Zhang et al., 2016).

The formation of piedmont glaciers is rare in the Eastern Mediterranean and likely uncommon in similar mid-latitude locations worldwide (Çiner et al., 2015) (e.g., Çiner et al., 2019; Hughes et al., 2006; Smith et al., 1997). Several hummocky moraine fields were reported by various researchers on glaciated mountains of Mediterranean mountains (e.g., Bayrakdar, 2012; Bayrakdar et al., 2017; Çiner, 2004; Çiner et al., 1999, 2017; Çiner & Sarıkaya, 2017; Cowton et al., 2009; Đurović, 2020; Köse et al., 2019, 2021, 2022; Moulin et al., 2022; Sarıkaya et al., 2009, 2014, 2017; Serrano-Cañadas & Martín-Moreno, 2018). Some of these mountains include hummocky moraine exposure ages derived from terrestrial cosmogenic nuclide dating (TCND) such as Mt. Erciyes (Sarıkaya et al., 2009), Akdağ (Sarıkaya et al., 2014), Geyikdağ

(Çiner et al., 2015; Sarıkaya et al., 2017), Bolkar Mountains (Çiner & Sarıkaya, 2017) and Mt. Dedegöl (Köse et al., 2019).

Several dating studies aimed to reconstruct and analyze the behavior of paleoglaciers in cirques and glacial valleys located in the Turkish mountains. Thanks to the TCND method, especially in the last two decades, a comprehensive glacial chronology for Türkiye has been established. Accordingly, the earliest glacial retreat in Türkiye is traced back to late Marine Isotope Stage 5 (MIS 5; 78.5 ± 17.6 ka) (Altınay et al., 2022). Nevertheless, the most recent ages, Late Holocene glacial deposits (1.2 ± 0.3 ka, 2.5 ± 0.3 ka, and 3.8 ± 0.4 ka), linked to the Neoglacial expansion, have also been identified (Sarıkaya et al., 2009). The local LGM (21.5 to 18.5 ka) in Quaternary glaciation areas in Türkiye is correlated with the global LGM (26.5 to 19 ka) (Sarıkaya & Çiner, 2015). Active glaciers in Türkiye are relatively scarce compared to the Quaternary glaciations. The total area of Turkish glaciers amounts to 12.29 km^2 , with Mt. Ararat (5137 m a.s.l.; the highest mountain of Türkiye) being the predominant contributor, encompassing over 60% of the region (Azzoni et al., 2020).

While the hummocky moraines created by piedmont glaciers represent uncommon phenomena in the Mediterranean, it is not surprising that there is a limited body of research on this subject. A comprehensive study (Çiner et al., 2015) conducted on Geyikdağ (2877 m a.s.l., SW Türkiye), an important paleo-piedmont glaciation site and vast hummocky moraine areas were dated by cosmogenic ^{36}Cl . They discovered that the ages of hummocky moraines originating from piedmont glaciers range from 16.0 ± 3.1 ka to 19.1 ± 3.4 ka based on 10 moraine boulders. The weighted average of all samples (excluding outliers) indicates that the earliest occurrence of glaciation took place approximately 18.0 ± 1.1 ka on Geyikdağ.

Within this study, we present a modest yet consistent glacial chronology of Mt. Davraz, while also providing an exploration of its comprehensive geomorphological features. Our analysis involved detailed mapping and topographical analyses of Mt. Davraz. We employed the terrestrial cosmogenic ^{36}Cl dating method to unveil the timing of the last deglaciation on Mt. Davraz. As our discussion unfolds, we delve into various dimensions, including assessing erosion rates on carbonate rock/erosion rates on TCND of carbonate rocks, interpreting deglaciation-age relationship, and uncovering the development of the Davraz hummocky moraine field. We also examined the potential influence of wind-induced snow redistribution on glaciation.

Finally, comparisons are drawn with glacial records in the nearby region, and intriguing insights are gained into piedmont glaciations and hummocky moraines within the broader Mediterranean context.

4.1.1 Study area

Mt. Davraz (37° 45' 16" N, 30° 43' 30" E; 2637 m a.s.l., above sea level) is located in the southwest of Türkiye (Figure 4.1a), between the city of Isparta and Lake Eğirdir (916 m a.s.l.). It is a part of the western section of the Taurus Mountain Range and appears like a stairway structure from north to south, cascaded by SW–NE-oriented active normal faults (Emre et al., 2011) (Figure 4.1b). Old thrust faults such as Antalya complex thrust (Paleocene) on the northeast and Lycian complex thrust (Langhian) with Aksu thrust (Late Miocene–Pliocene) on the southeast set the boundaries of Mt. Davraz (Poisson et al., 2003). It has also been argued that Mt. Davraz represents a slice of the Beydağları Massif (a shallow-marine carbonate platform) by Poisson et al. (2003). The lithology of Mt. Davraz is mostly dominated by Triassic and Jurassic–Cretaceous neritic limestones.

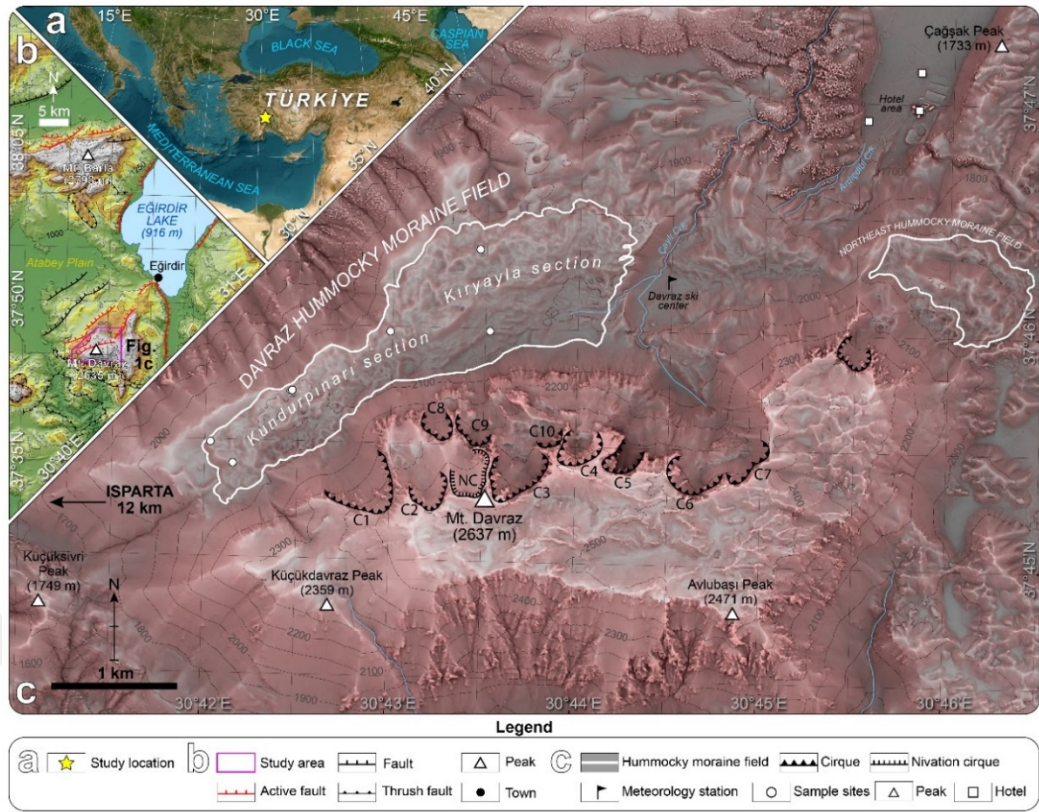


Figure 4.1 : Location maps of the study area: (a) location of the study area within Türkiye, (b) showing the study area and its immediate surroundings, along with their structural features, on the digital elevation model (DEM) and (c) a red relief image map (RRIM, Chiba et al., 2008) showing the location of the study area, Davraz hummocky moraine field, northern slopes of Mt. Davraz (C#: cirques names; NC: nivation cirque).

The climate of Türkiye is regulated by seasonal changes between mid-latitude and sub-tropical/tropical pressure and wind systems, resulting in a subtropical Mediterranean macroclimate with hot and dry summers (Lolis & Türkeş, 2016). The distribution of Mediterranean moisture across Anatolia is significantly influenced by the Taurus Mountains. This mountain range acts as a natural barrier between the coastal and interior regions due to their high altitudes (>3 km, a.s.l) (Kuzucuoglu & Roberts, 1997; Sarıkaya et al., 2009). The study area is located on a transition between coastal areas and interior regions. It is characterized by a sub-humid climate (Akman & Ketenoğlu, 1986; Iyigün et al., 2013). The average summer and winter temperatures in the study area are 16.6 °C and -0.9 °C, respectively, while the mean annual temperature is 7.8 °C, based on data measured from Isparta Davraz Ski Center weather station (37° 46' 13" N, 30° 44' 32"E; 1951 m a.s.l., however, only 5 years of observation, since 2017).

4.1.2 Previous studies

The first glacial study on Mt. Davraz was carried out by Louis (1944) within the context of Pleistocene glaciations in Anatolia. He identified several cirque glaciations on the mountain. The most detailed research was done by Ardos (1977b), who focused on the glacial geomorphology of the mountain. He mapped the development of cirques, moraines, and rock glaciers on the northern slopes of Mt. Davraz. Ardos (1977b) also remarked the equilibrium line altitude (ELA), based on the cirque floor altitudes, as in between 2400 and 2450 m a.s.l. during Pleistocene.

A recent regional study was done by Evans et al. (2021) about the paleoclimatic evaluations from morphometric features of the cirques of all Western Taurus Mountains of Türkiye. Ten cirques of Mt. Davraz were part of this study. Based on their results, glacier mass balance at times of cirque development is mainly controlled azimuthally by solar radiation contrasts between shaded and exposed slopes (Derbyshire & Evans, 1976; Evans, 1977). However, this conclusion is not supported by the data from Mt. Davraz. Because the vector means aspect value of the cirques on Mt. Davraz ($347^{\circ} \pm 17^{\circ}$) does not agree with the overall vector mean aspect ($015^{\circ} \pm 12.5^{\circ}$ - confidence limits: 002° and 027°) of the cirques in the Western Taurus. They also calculated the LGM ELA of Mt. Davraz as 2233 m a.s.l, using moraine-based glacier reconstructions, independent of cirque floors' altitude.

4.2 Methodology and Methods

4.2.1 GIS and map producing

Mapping was carried out utilizing 1:25.000 scale topographic maps provided by The General Directorate of Mapping of Türkiye (HGM; *Tr. Harita Genel Müdürlüğü*), high-resolution satellite imagery from sources such as Bing Maps, Google Earth, and HGM-Küre (<https://kure.harita.gov.tr/>), as well as extensive fieldwork conducted during the summer of 2019 and autumn of 2020. During the field studies, oblique images were captured using a UAV in the study area. These images were subsequently utilized for interpreting the geomorphological features of the region. Digital elevation/surface models (DEMs/DSMs) with pixel resolutions of 30 m (SRTM 1 arc-second global, obtained from <https://earthexplorer.usgs.gov/>) and 5 m (generated from HGM stereo aerial photos) were utilized in the generated maps, depending on the

scales of the maps. DSMs (with a 5-m pixel resolution) were utilized to generate cross-sections illustrating structural features, such as faults and depressions. To enhance the visualization of the topography, both a Hillshade and a Red relief image map (RRIM, Chiba et al., 2008) were generated from the DEM. We calculated the LGM ELA using a GIS toolbox developed by Pellitero et al. (2015). The Area-Altitude Balance Ratio (AABR) method was employed for the calculation of the LGM ELA on Mt. Davraz. The coordinate reference system chosen for the study area was WGS 84/UTM zone 36 N EPSG: 32636.

4.2.2 Terrestrial cosmogenic nuclide (TCND)

The cosmic ray exposure ages of moraines in Mt. Davraz were determined by using in situ-produced cosmogenic ^{36}Cl . We collected six carbonate rock samples from the boulders of the hummocky moraine field (Table 4.1). While four of the samples were extracted from boulders located on crests near the external boundaries, the two last samples were collected on boulders from the inner part of the area (Figures 4.1c and 4.2a-f).

Table 4.1 : Geographic sample locations, boulder dimensions, sample thicknesses, and topographic shielding factors of the cosmogenic ^{36}Cl samples. Attributes and local corrections to production rates for sampled Davraz hummocky moraine boulders.

Sample ID	GPS Latitude (WGS84) °N (DD)	GPS Longitude (WGS84) °E (DD)	Elevation from the topographic map (m)	Boulder dimensions (LxWxH) (m)	Sampled thickness (cm)	Topography shielding factor (-)
DVRZ19-01	37.76198	30.70778	2096	2 x 3.7 x 1.8	3	0.99484
DVRZ19-02	37.76622	30.71650	2086	2.4 x 1.1 x 0.7	1	0.99859
DVRZ19-03	37.77230	30.72457	2047	1.2 x 0.7 x 0.7	1	0.99812
DVRZ20-04	37.75675	30.70233	2068	1.1 x 0.9 x 0.3	0.7	0.98462
DVRZ20-06	37.75822	30.70057	2058	0.9 x 0.9 x 0.4	2.5	0.99477
DVRZ20-08	37.76658	30.72553	2032	2.3 x 0.9 x 0.7	1	0.95519

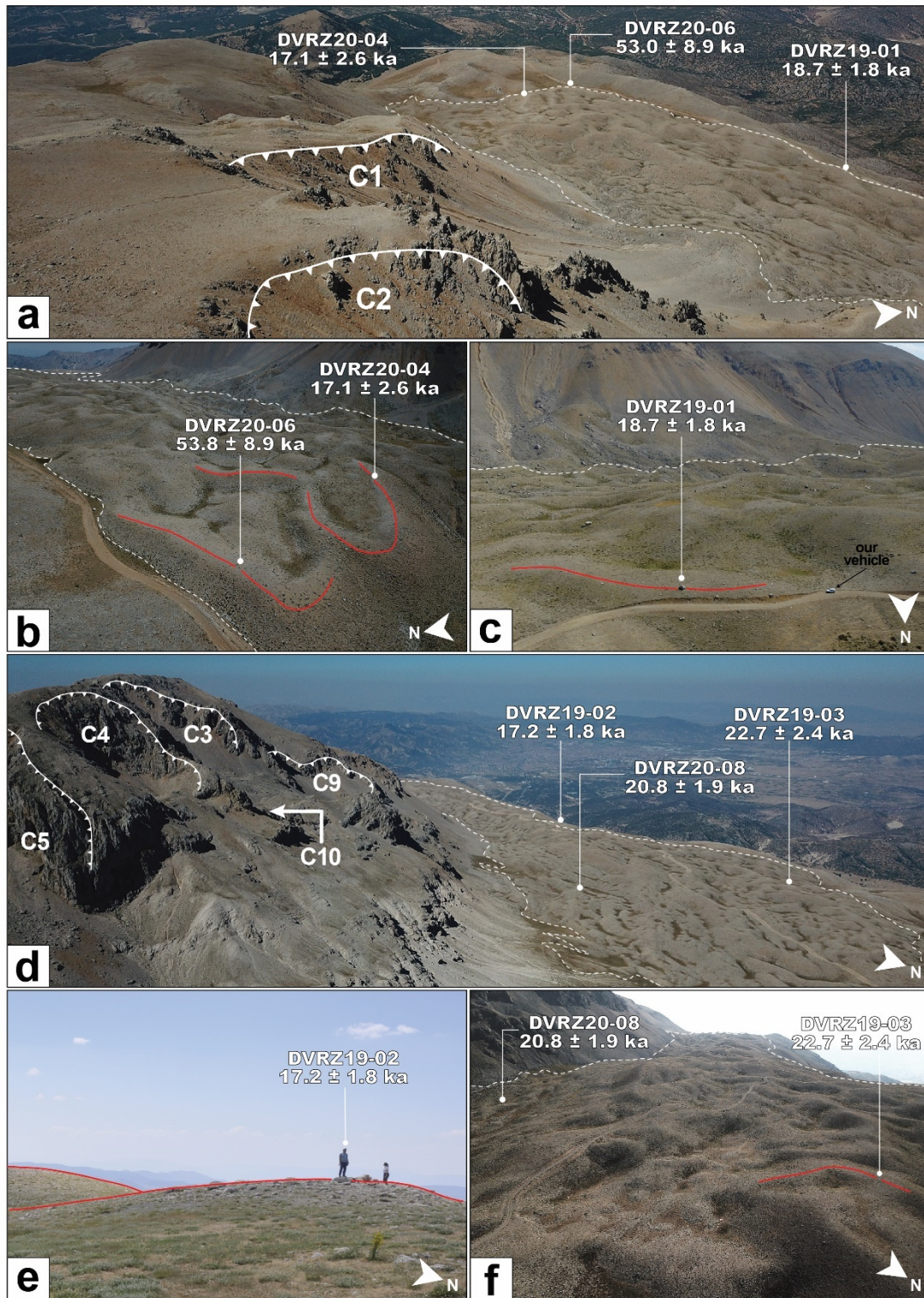


Figure 4.2 : Locations of cosmogenic samples on the images captured from the UAV camera (excluding 'e', which was taken with a Nikon D90 SLR). The hummocky moraine borders are depicted by dashed white lines, the cirques are represented by spiky white lines, and the moraine crests are indicated by red lines.

The coordinates of sampled boulders were recorded using a handheld GPS device. To ensure optimal exposure to cosmic rays, preference was given to flat or gently sloping

moraine crests. Horizon angles to be used in topographic shielding calculations were measured at each sampling site using a compass and a clinometer at 45° azimuthal intervals. A hammer and chisel were used to collect rock samples from boulder tops on the crests of the moraines (Figure 4.3). Our selection of boulders was determined by their size, position, and post-depositional preservation.

The chemical preparations were followed by the procedures described by Sarıkaya (2009). Preparation of the samples was done at ITU-Kozmo-Lab (Istanbul Technical University, Türkiye) (<https://kozmo-lab.itu.edu.tr/en>). The carbonate rock samples were crushed and ground to a fractional size of 0.25 mm to 1 mm before being analyzed. Samples were leached with milli-Q and dilute nitric acid (10%) to remove meteoric chlorine and biogenic material. Then, natural chlorine in the rock matrix was liberated by dissolving leached rocks in HDPE bottles within the solution of 400 ml 10% HNO₃. Before complete dissolving, the materials were mixed with natural NaCl (Merc Emsure) and ³⁵Cl (99.7%) enhanced carrier (Aldrich). By adding AgNO₃, chlorine was precipitated. By precipitating the samples twice as BaSO₄, ³⁶S, and an isobar of ³⁶Cl, the rate in the detector can be kept low enough not to influence the identification of ³⁶Cl (Mechernich et al., 2019). The Australian Nuclear Science and Technology Organization (ANSTO, Sydney-Australia), analysed chlorine isotopic ratios of the samples, using its 6 MV SIRIUS Tandem Accelerator (Wilcken et al., 2017, 2019).



Figure 4.3 : Moraine blocks sampled in the Davraz hummocky moraine field. Sampled surfaces are indicated by white arrows.

Purdue Z93-0005 ($^{36}\text{Cl}/\text{Cl}$: 1.2 ± 10^{-12} with natural $^{37}\text{Cl}/^{35}\text{Cl}$ ratio) and KN, provided by Dr. Kunihiro Nishiizumi (1.6 ± 10^{-12} and 5.0 ± 10^{-13} , $^{36}\text{Cl}/\text{Cl}$ ratios; Sharma et al., 1990), respectively, serve as the major and secondary AMS standards (Wilcken et al., 2013). A ^{36}Cl half-life of 3.014×10^5 years was calculated using the decay constant of $2.303 \pm 0.016 \times 10^{-6} \text{ year}^{-1}$. Utilizing the isotope dilution technique, total Cl was calculated (Wilcken et al., 2013). The concentrations of all major and particular trace elements in the samples were analyzed at Activation Laboratories, Inc. in Ontario, Canada (Table 4.2). Average CaO concentrations are 54.95%, although K_2O concentrations are very low and typically close to the detection limits (0.01%), except in Sample #2 (0.07%, still very low) (starting from this point, we will use “#” instead of the last two digits for sample identification; e.g., sample #1 instead of DVRZ19-01). The natural Cl concentrations are also very low, between 3.6 and 24.1 ppm. Thus,

the main production mechanism for ^{36}Cl is from the spallation of ^{40}Ca . For all samples, it was assumed that the rock density was 2.6 g/cm^3 . Table S1 contains all the information needed to recalculate the age of samples and natural ^{36}Cl concentrations.

The sample ages were determined using the CRONUS Web Calculator version 2.1 (<https://cronus.cosmogenicnuclides.rocks/2.1/>) (Marrero et al., 2016b). Corrections for thickness and topographic shielding were done for all production rate calculations. We also used a snow correction (~ 0.97) on spallation responses with 50 cm snow thickness on top of moraine boulders, assuming a snow cover period from December to April. We utilized the corrected ^{36}Cl production rates reported by Marrero et al. (2021) [$55.9 \pm 4.5 \text{ atoms } ^{36}\text{Cl} (\text{g Ca})^{-1} \text{ a}^{-1}$ for Ca spallation, $153 \pm 12 \text{ atoms } ^{36}\text{Cl} (\text{g K})^{-1} \text{ a}^{-1}$ K spallation and $696 \pm 196 \text{ fast neutrons } (\text{g air})^{-1} \text{ a}^{-1}$]. The Fe and Ti spallation production rates of ^{36}Cl are $1.9 \text{ atoms } (\text{g Fe})^{-1} \text{ a}^{-1}$ and $3.8 \text{ atoms } (\text{g Ti})^{-1} \text{ a}^{-1}$, respectively (Marrero et al., 2016a). The time-dependent Lifton-Sato-Dunai method (also called ‘LSD’ or ‘SF’ scaling) (Lifton et al., 2014), was used to scale ^{36}Cl production rates.

Table 4.2 : Chemical compositions of the bulk rock samples after leaching with dilute nitric acid (10%). Analysis performed at the AcmeLab, Canada (Bureau Veritas Commodities Canada Ltd.) by ICP-ES (major elements), and ICP-MS (trace element).

Sample ID	Major Elements										Trace Elements					
	Al ₂ O ₃ (wt%)	CaO (wt%)	Fe ₂ O ₃ (wt%)	K ₂ O (wt%)	MgO (wt%)	MnO (wt%)	Na ₂ O (wt%)	P ₂ O ₅ (wt%)	SiO ₂ (wt%)	TiO ₂ (wt%)	CO ₂ (LOI) (wt%)	Sm (ppm)	Gd (ppm)	U (ppm)	Th (ppm)	Cl (ppm)
DVRZ19-01	0.01	55.65	<0.04	<0.01	0.41	<0.01	0.01	0.08	0.09	<0.01	43.70	0.07	0.15	2.20	<0.2	16.3 ± 1.9
DVRZ19-02	0.25	54.59	0.18	0.07	0.44	0.03	0.02	0.04	1.14	0.01	43.10	0.58	0.93	0.20	<0.2	3.6 ± 0.3
DVRZ19-03	0.16	54.93	0.20	0.02	0.59	<0.01	0.03	0.01	0.45	<0.01	43.60	<0.05	0.08	6.20	<0.2	24.1 ± 1.6
DVRZ20-04	0.03	54.81	0.28	0.01	0.33	<0.01	0.01	0.05	0.57	<0.01	43.90	0.15	0.28	1.90	<0.2	16.0 ± 0.3
DVRZ20-06	0.04	54.99	0.24	<0.01	0.24	<0.01	<0.01	0.05	0.25	<0.01	44.10	0.13	0.29	2.20	<0.2	8.4 ± 0.5
DVRZ20-08	0.03	54.73	0.37	<0.01	0.44	<0.01	<0.01	0.02	0.26	<0.01	44.10	<0.05	0.07	4.50	<0.2	11.5 ± 0.6

We calculated the ages with and without accounting for erosion. We employed the erosion rates on carbonate rock rates of Mt. Barla and Mt. Dedegöl from Hashemi et al. (2023), due to their proximity to Mt. Davraz. The rates are 11.72 ± 1.17 (at 1630 m a.s.l.) and $12.49 \pm 1.25 \text{ mm ka}^{-1}$ (at 1935 m a.s.l.). Therefore, we assumed an erosion rate of 10 mm ka^{-1} for Mt. Davraz.

Based on Griffin et al. (2022), to integrate age estimates from multiple TCND ages for each surface, we consider the uncertainty of the age of the surface as normally distributed. Employing Monte Carlo methods, we randomly sample 10000 times from the age distribution of each representative TCND age, combining these samples to derive the mean and standard deviation for the estimated age of the surface. Landform

ages were calculated using a weighted mean. The analytical and production rate uncertainties were included in the age calculations with a 1-sigma uncertainty for all ages.

4.3 Results

4.3.1 Glacial geomorphology

There are no active glaciers on Mt. Davraz today. However, the geomorphology of the mountain provides evidence of glaciation in the past. On the northern slopes of the mountain, there exists a relatively extensive moraine area in comparison to the overall coverage of Mt. Davraz, accompanied by successive cirques developed on the upper slopes. It can be asserted that the magnitude of these glaciations is greater on the northern slope than on the southern and eastern slopes of the mountain. The Davraz hummocky moraine field on the north and cirque-dominated slopes on the northern slopes of the mountain are the two main parts of glacial geomorphology in the study area (Figure 4.4). We calculated the ELA during the LGM period as 2161 m a.s.l. For the Western Taurus region, including Mt. Davraz, the modern ELA depression has been indicated by Akçar (2022) to be between 1000 and 1400 m a.s.l.

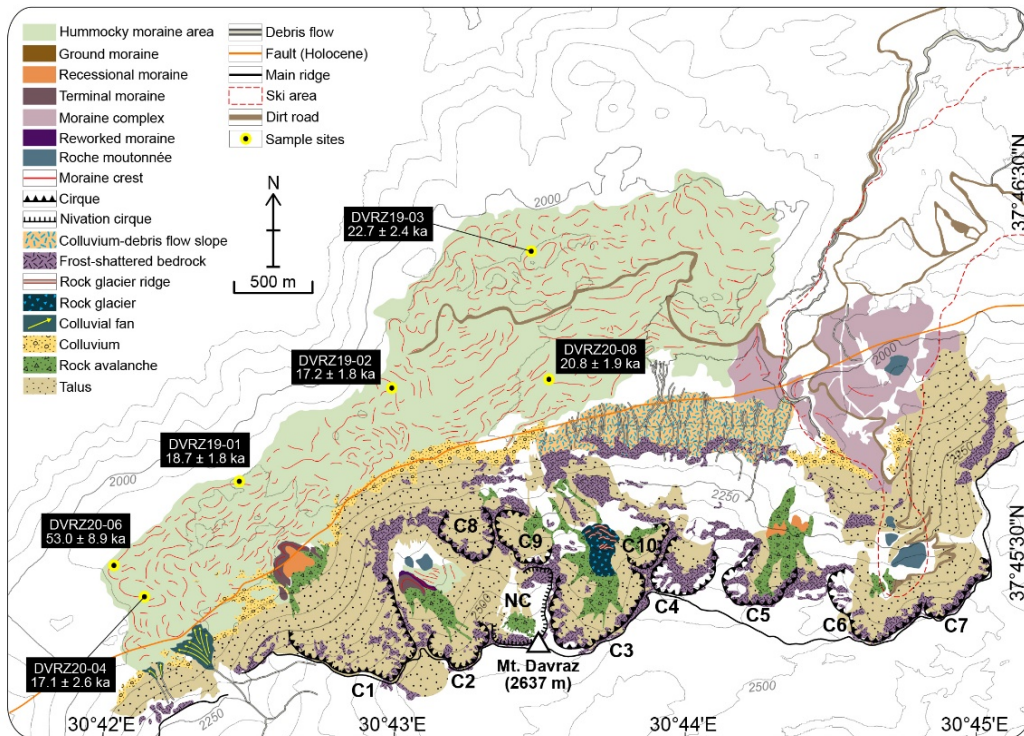


Figure 4.4 : The geomorphological map of the northern slopes of Mt. Davraz (C#: cirques names; NC: nivation cirque).

Mid-latitude glaciated mountains often show signs of previous glaciation as cirque development. These sites were characterized by a particular topography that controlled the accumulation and movement of ice, with ice flowing from the back walls of cirques towards the margins of glaciers over short distances. Ardos (1977b) mentioned three large cirques and a few minor nivation hollows on Mt. Davraz. However, he mentioned neither the formation of the hummocky moraines nor any piedmont glacier formation in his study. We mapped 10 cirques and a nivation hollow on the northern slope of Mt. Davraz unlike Ardos (1977b) (Figures 4.1c and 4.4).

Topographical restrictions prevented the cirques from forming proper glacial valleys. However, paleoglaciers developed in these cirques have been able to form piedmont glaciers despite the absence of well-developed glacial valleys. Accordingly, it is possible to say that cirque glaciers directly started to accumulate on the slopes and then slowed into the flat-lying area located in front of the northern slope of Mt. Davraz (Figure 4.1c). Hereby, we observed evidence of paleo-piedmont glaciers developed by these cirque glaciers during the LGM (c. 26.5 to 19 ka). As the late LGM marked the onset of deglaciation, piedmont glaciers began to diminish and retreat. Throughout the retreat, spanning from the late LGM to the Late-glacial (c. 19 to 11.7 ka; cf. Penck & Brückner, 1901/1909; Brazier et al., 1998; Böhlert et al., 2011; Shrestha & Aryal, 2011; Hagg, 2022) period, a substantial quantity of hummocky moraines was deposited on these relatively flat surfaces. As mentioned above, piedmont glaciers may also originated from ice caps extending on mountain tops, instead of being fed by cirque glaciers. Nonetheless, during our field surveys, we could not find evidence of ice cap formation on the top of Mt. Davraz.

Besides the northern Davraz hummocky moraine field, there is another hummocky moraine field located on the eastern side of the mountain, which we call the Northeast hummocky moraine field (Figure 4.1c). Presumably, it was formed by a single cirque glacier. There, a part of this hummocky moraine field proximal to the slope was reworked by a rock glacier. The rock glacier was probably formed by frost-shattering processes on a well-developed talus slope. However, we preferred to exclude these two areas from this study and focused on the largest ‘Davraz’ hummocky moraine field to ensure the integrity of the research.

The glacial geomorphology of Mt. Davraz changed with the onset of postglacial modification. The relatively thick talus slopes that developed on the walls of the

cirques are the most important of these changes in the geomorphology of Mt. Davraz. These slopes were developed by the combined impact of paraglacial and periglacial processes that dominated after the deglaciation. Talus slopes are generally seen on the steep slopes of the mountain apart from the cirque headwalls. Especially in the areas where these cirque headwalls are located, rock falls due to paraglacial and periglacial processes are common. Finally, limited fluvial processes have continued to be effective on the mountain. They caused the formation of gully channels and debris flows. Colluvial fans and deposits were formed in certain areas due to these processes.

The Davraz hummocky moraine field covers a larger part of the northern slope of Mt. Davraz, as introduced above. It extends approximately 4 km from southwest to northeast, covering an area of about 4 km². On the western margin of this area, moraines show slightly terminal moraine morphology (Figure 4.1c). These moraines can be considered as the terminus part of the Davraz paleoglaciers. Contrary to the western part, moraines on the eastern margin of the area are difficult to distinguish, especially from the karst morphology due to postglacial karstification and fluvial incision. Thus, we divided the area into two sections to a better definition of the hummocky field as Kundurpinarı and Kırıyayla sections (Figure 4.1c). In making this division, we have used some of the notable differences between these two sections. For instance, Kırıyayla has a larger area than the Kundurpinarı section. Concerning that, the hummocky moraine field shows a larger area from the west (Kundurpinarı) to the east (Kırıyayla). It also loses elevation, roughly 60 m, along the overall length of the field. At the eastern terminus part, hummocky moraines intertwined with karst morphology. The moraine boulder of both sections usually consists of poorly sorted limestone clasts in a fine-grained matrix with boulder-to-debris size (Figure 4.5).



Figure 4.5 : Moraine outcrops from the Davraz hummocky moraine field.

Geomorphological mapping and oblique images (derived from UAV and Google Earth) show that the moraine crests vary in distribution and morphometry. While proximal moraine lobes have relatively small scales, distal moraines in the Kundurpınarı section are larger and well-developed. Towards the east, the Kundurpınarı hummocky moraine field becomes narrower, and then the Kırıyayla section starts. Here, the slope-proximal area has a lower relief compared with the whole hummocky field. The northern and eastern boundary of this area is partly removed by fluvial processes. Based on debris flow developments, the northern slopes of the moraines intertwined with debris flow valley heads. Therefore, it has become difficult to determine the boundaries of the moraines in this area.

4.3.2 Glacial chronology

All samples were collected from the tops of boulders situated on the crests of hummocky moraines (Figures 4.1c and 4.2 and Table 4.3). Four of them were from the Kundurpınarı section while two of them were taken from the Kırıyayla section. Five of these ages were internally consistent (#1: 18.7 ± 1.8 ka, #2: 17.2 ± 1.8 ka, #3: 22.7 ± 2.4 ka, #4: 17.1 ± 2.6 ka, #8: 20.8 ± 1.9 ka). However, one of them (#6: 53.0 ± 8.9

ka) is a few standard deviations older than the others; thus, we considered this age as an outlier.

Table 4.3 : Cosmogenic ^{36}Cl inventories, boulder ages, and landform ages of the Davraz hummocky moraine field. The ages calculated without erosion are accompanied by internal uncertainties provided in brackets. Cosmogenic ^{36}Cl inventories, boulder ages, and landform ages of the Davraz hummocky moraine field. TCND outliers are indicated with an asterisk (*) at the end of the sample IDs.

Sample ID	³⁶ Cl age			
		without erosion	erosion corrected	Landform
	³⁶ Cl (measured)	correction	(10 mm ka ⁻¹)	age
	(10 ⁴ atoms g ⁻¹ rock)	(ka)	(ka)	(ka)
Kundurpinarı Section				
DVRZ19-01	148.9 ± 9.5	16.7 ± 1.5	18.7 ± 1.8	17.7 ± 2.2
DVRZ19-02	128.6 ± 7.8	14.9 ± 1.3	17.2 ± 1.8	
DVRZ20-04	131.1 ± 16.4	15.4 ± 2.1	17.1 ± 2.6	
DVRZ20-06*	329.1 ± 18.1	35.8 ± 3.1	53.0 ± 8.9	
Kıryayla Section				
DVRZ19-03	179.9 ± 10.1	19.9 ± 1.5	22.7 ± 2.4	21.8 ± 2.4
DVRZ20-08	142.8 ± 7.6	18.1 ± 1.5	20.8 ± 1.9	
Average=				19.3 ± 3.0

We obtained three landform ages based on two scenarios: (1) a uniform glacier retreat with a single mean landform age (Figure 4.6a and Table 4.3); and (2) a gradual glacier retreat with two mean landform ages (Figure 4.6b,c and Table 4.3). We adopted the methodology of Griffin et al. (2022) to estimate the age of landforms, despite the larger uncertainty in the approach. However, we consider that larger uncertainties better represent the landform age, especially if there is a limited age dataset.

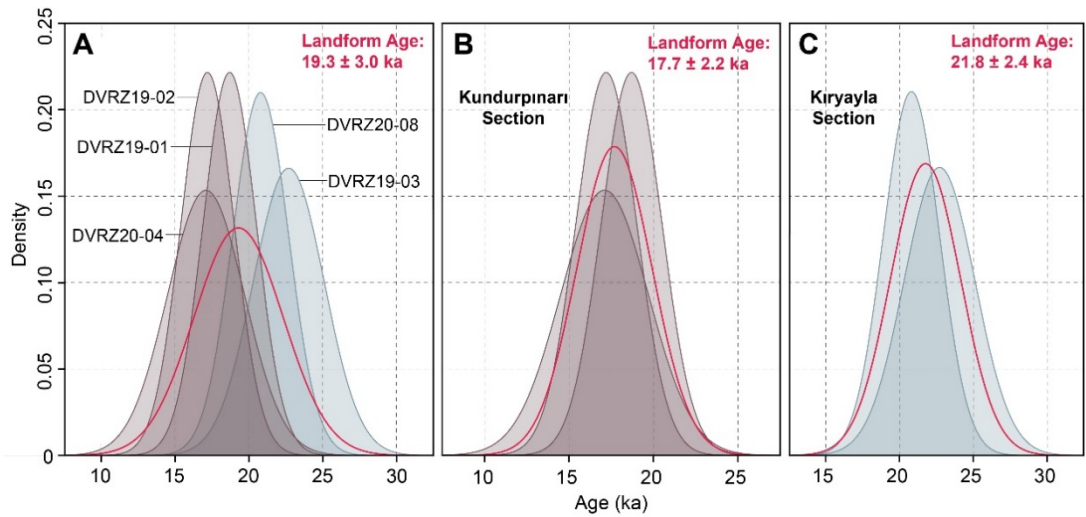


Figure 4.6 : Diagrams depicting the distribution of the individual cosmogenic ^{36}Cl surface exposure ages ($n = 5$, outliers excluded). (a) the mean surface age was obtained by combining all ages calculated according to the first glacier retreat scenario; surface ages were calculated separately for the Kundurpınarı (b) and Kırıyayla (c) sections, following the second glacier retreat scenario.

4.4 Discussion

4.4.1 Denudation/erosion rates on carbonate rocks

Rates of denudation in karst landscapes, encompassing both chemical denudation and mechanical erosion processes (Ford & Williams, 2007), can exhibit substantial variations. The preservation of glacial deposits in glaciokarst landscapes is facilitated by efficient rainfall percolation, limited fluvial reworking, and the cementation of landforms (Woodward et al., 2008; Žebre & Stepišnik, 2015; Çiner et al., 2015). In a recent study, Allard et al. (2020) investigated the scattered ages of moraines within the Tymphi massif in the Pindus Mountains, northwest Greece. They correlated these findings with moraine degradation, attributing it to the erosion of limestone surfaces, which is more prevalent in glaciokarst landscapes compared to granitic or siliceous lithologies. This phenomenon arises from the presence of karst depressions and the solubility of limestone, as highlighted by Putkonen & Swanson (2003). Despite the limited number of samples collected from our study area, we believe that we have obtained well-clustered age results. Consequently, we do not anticipate any issues arising from moraine degradation.

In their erosion corrections applied to ages calculated using the ^{36}Cl isotope on carbonate rocks, Žebre et al. (2019) assumed an erosion rate of 40 mm ka^{-1} (with the

upper limit calculated even at 60 mm ka^{-1}). This choice was influenced by the fact that the study area is one of the regions (The Dinaric Mountains) with the highest rainfall in Europe. On the other hand, cosmogenic dating investigations focused on carbonate lithology in the Taurus Mountains, researchers have adopted an erosion rate of either 5 mm ka^{-1} (Sarıkaya et al., 2014; Çiner et al., 2015; Çiner & Sarıkaya, 2017; Köse et al., 2022) or 10 mm ka^{-1} (Sarıkaya et al., 2017; Çiner et al., 2017; Köse et al., 2019; Altınay et al., 2022). In a recent study by Hashemi et al. (2023), the long-term denudation rate of carbonate rocks in the Taurus Mountains was determined to range from 1.92 ± 0.31 to $45.77 \pm 3.91 \text{ mm ka}^{-1}$ using the cosmogenic ^{36}Cl isotope.

In certain investigations, due regard was given to the depth of the solution grooves (cf., rinnenkarren; Bögli, 1980) when ascertaining the erosion rate. In certain investigations, careful consideration was given to the depth of the solution grooves when determining the erosion rate. For instance, in studies conducted in the Taurus Mountains, solution grooves ranging from a few centimeters to 10-15 centimeters (Sarıkaya et al., 2017; Çiner et al., 2017) were inferred as 10 mm ka^{-1} erosion. On the other hand, in areas with solution grooves measuring several centimeters, the erosion rate was determined to be 40 mm ka^{-1} for Velež and Crvanj mountains and Blindje polje in Bosnia and Herzegovina (Žebre et al., 2019; Çiner et al., 2019). In our study area, the maximum depth of solution grooves on moraine boulder surfaces is only a few centimeters, contrasting with other studies that assume a 10 mm ka^{-1} erosion rate. In karst terrains, extreme cold climatic conditions lead to a scarcity of liquid water, limiting dissolution and allowing other geomorphological processes to dominate morphological evolution (Ford & Williams, 2007). Given the interior location (i.e., transition to continental climate) of Mt. Davraz, the extreme cold may explain why we could not identify relatively deeper solution grooves on the sampled moraine boulders. Chert veins can also serve as indicators to estimate differential weathering rates on limestones. Allard et al. (2020) used the thicknesses of elevated chert veins in limestone to assess the varying erosion rates of glacial boulders in Greece. While they proposed cosmogenic ^{36}Cl ages corrected for erosion as low as 0.6 mm ka^{-1} , they acknowledge the possibility of higher rates (greater than 10 mm ka^{-1}) within their study area in Greece. In another study conducted by Köse et al. (2022) in the eastern part of the Central Taurus Mountains, the erosion rate was determined as 3 mm ka^{-1} based on the chert vein thickness. This rate was evaluated alongside the erosion rate of 7 mm

ka^{-1} obtained from a depth profile of an alluvial fan (Sarıkaya et al., 2015a) near the region, and the erosion rate on the surface of the moraine blocks was accepted as 5 mm ka^{-1} . However, we were unable to locate any chert veins on moraine boulders in our study area.

Considering the previous cosmogenic ^{36}Cl dating studies and the published denudation rates by Hashemi et al., (2023) for the Taurus Mountains, we have decided to apply an erosion rate of 10 mm ka^{-1} for the ages obtained from the moraine blocks in the case of Mt. Davraz. Consequently, the cosmogenic ages we obtained are consistent with the local LGM (21.5 to 18.5 ka), indicating that the erosion rate of 10 mm ka^{-1} is appropriate.

4.4.2 Evolution and age interpretation of the hummocky moraine field

Based on our morphological observations, we are assuming that the hummocky moraine field preceding the glacier occupation was characterized by relatively flat terrain, resembling the plateau topography observed at the summit of Mt. Davraz (Figure 4.1c). Accordingly, the Kırıyayla section appears to exhibit lower relief compared to the Kundurpınarı section after the disappearance of the Davraz paleoglacier (Figure 4.7). We inferred that the more widely expanded paleoglacier in the Kırıyayla section is the direct cause of this difference. Because, as the glacier expands, its thickness tends to decrease, especially in the ablation area (Flint & Demorest, 1942). Correspondingly, thinner glaciers will have less debris to carry vertically, so the morphology of the hummocky moraines formed by ice wastage may be relatively low. On the other hand, the Kundurpınarı section has a higher relief. We believe that this can be attributed to the longitudinal structure of this section. As the paleoglacier did not advance significantly in this area, it might have expanded without losing much of its maximum thickness. The debris carried by the glacier vertically might have formed higher moraines here because it was deposited in a relatively narrow region.

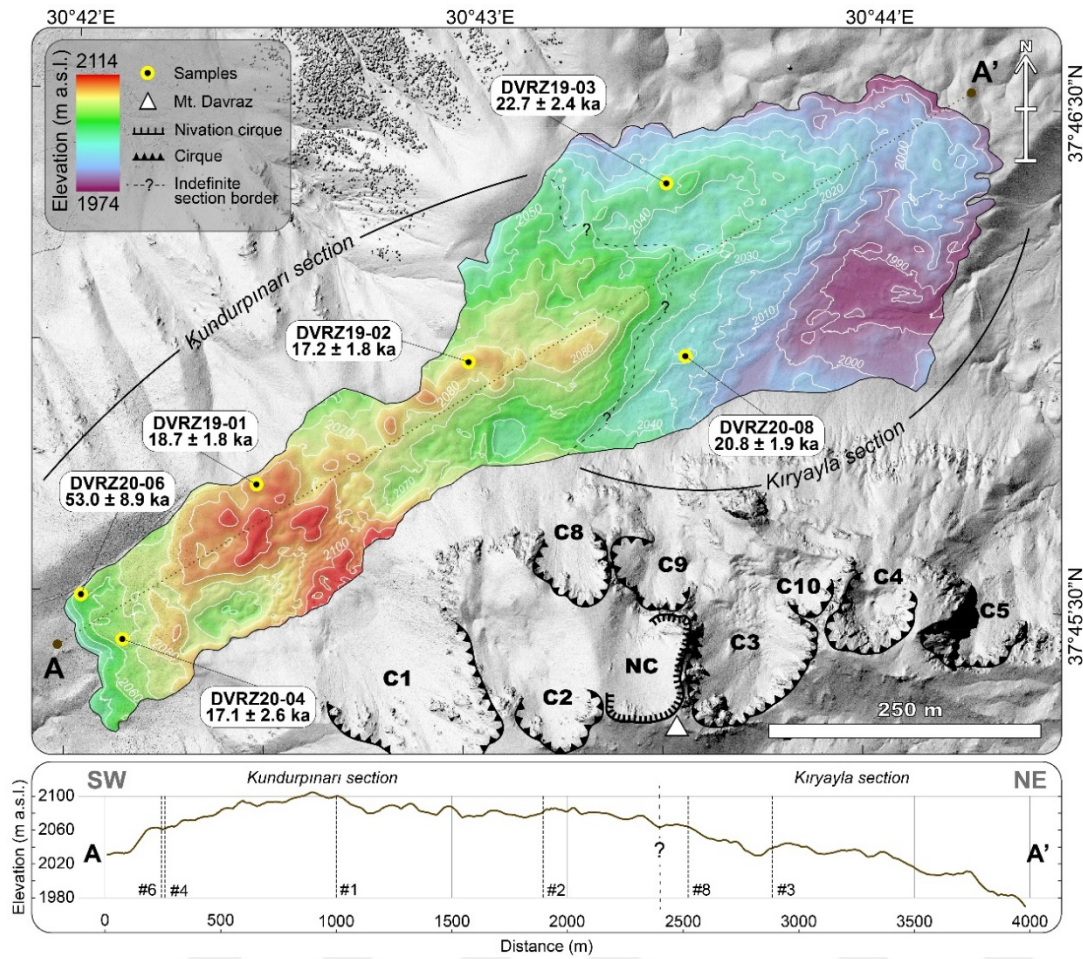


Figure 4.7 : The Davraz hummocky moraine field is shown on the DSM with field sections, cosmogenic ages, and cirques (top), while the graph (bottom) displays the southwest-northeast oriented profile across this field.

A multitude of factors, including the quantity of debris transported by ice, the location of the debris on, within, or beneath the ice, the quantity of meltwater, and the resulting erosion and deposition, can lead to the formation of diverse landforms (such as hummocky moraines) as a consequence of disintegration (Gravenor & Kupsch, 1959). Hummocks have a chaotic appearance due to their random distribution and varying dimensions. Therefore, it may be challenging to understand deglaciation patterns in hummocky moraine fields. However, the morphology and TCND ages of the Davraz hummocky moraine field enable us to deliberate on the pattern of glacier retreat in this area. At this point, we believe that the relatively slight variations in the clustered ages are also connected to moraine stabilization resulting from the melting of dead ice caused by ice disintegration. The distribution and morphology of moraines, combined with our cosmogenic ages, suggest that the deglaciation of the Davraz paleoglacier began in the Kıryayla section and terminated in the middle of the Kundurpinarı section

(Figure 4.8). We believe that the deglaciation in the area was led by the formation of four primary paleoglacier patches due to ice stagnation-dominated deglaciation. The paleoglacier patches centered with samples #1, 2, 3, and 4 (Figure 4.8). However, the moraine morphology shows that the slope-proximal part of the glacier retreated into the slope and disappeared in the Kırıyayla section (Figure 4.8b–d). In this respect, we do not consider there was an ice-stagnation-controlled melt-out occurred in this segment of the paleoglacier.

In the first deglaciation scenario, we considered all ages to ascertain the optimal representation of the geomorphological age. After excluding the outlier, Sample #6 (53.0 ± 8.9 ka), we calculated the landform age of the Davraz hummocky moraine field as 19.3 ± 3.0 ka (Figure 4.6a). In a recent study by Tielidze et al. (2022) in the Southern Alps, similar results were obtained for the ages represented in these two scenarios on a terminal moraine unit with a hummocky character. The landform age obtained from 11 boulders from the northern part of the above-mentioned moraine is 20.1 ± 0.4 ka, while on the southern part, it is calculated as 19.5 ± 0.4 ka. However, since this landform is considered as a single continuous moraine, its formation is interpreted as occurring 19.8 ± 0.3 ka. In this regard, the Davraz hummocky moraine bears similarity to the first scenario in terms of the deglaciation process. Alternatively, our results suggest a two-phased deglaciation at 21.8 ± 2.4 ka and a more recent phase at 17.7 ± 2.2 ka (Figure 4.6b,c) in the second scenario. The shared outcome of the scenarios is that the glaciers began their retreat during the global LGM (26.5 to 19 ka) and concluded it throughout the Late-glacial (c. 19 to 11.7 ka). However, considering the morphological evidence and the obtained ages, we posit a two-phase deglaciation, with the second scenario appearing more consistent. Hence, we find it appropriate to conduct the discussion based on the second scenario.

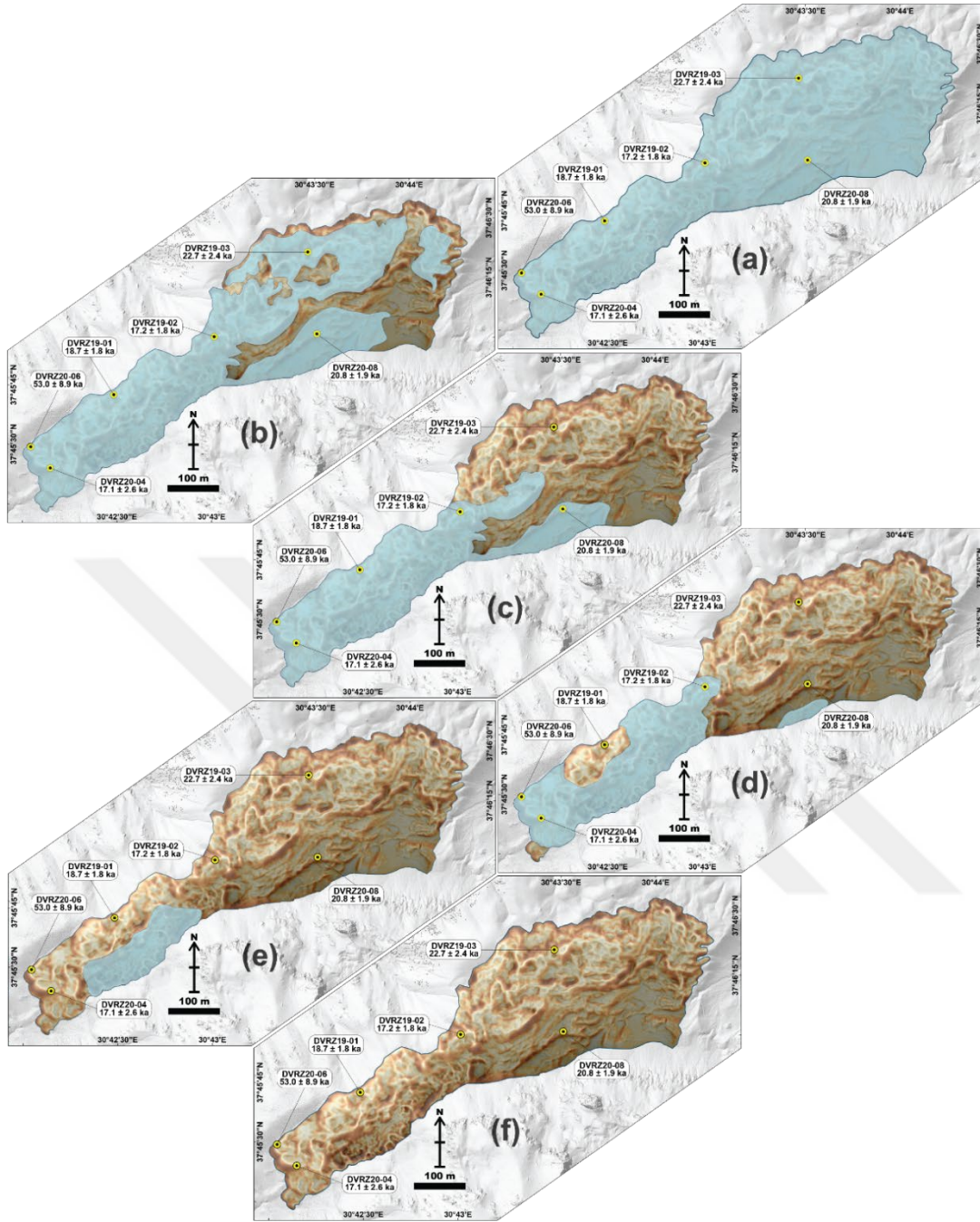


Figure 4.8 : Speculative representation of the melt-out/retreat of the Davraz paleoglacier, progressing from the east (Kırayla section) to the west (Kundurpinarı section) with cosmogenic ages.

We have two ages in the Kırayla section as Sample #3 (22.7 ± 2.4 ka; 2060 m a.s.l.) and #8 (20.8 ± 1.9 ka; 2038 m a.s.l.). Sample #3, the oldest age, is located close to the terminus of the glacier. By examining the location of this sample and its surroundings, one can observe the presence of relatively high moraines (Figure 4.7). This area was among the first parts to separate from the paleoglacier and commenced melting, following the onset of the stagnant ice period after the deglaciation of the Davraz

paleoglacier (Figure 4.8b,c). During the retreat and stagnation of the Davraz paleoglacier, the southwest boundary of the area is most likely controlled by the fault that marks this boundary (Figure 4.1b). This fault, the so-called Holocene fault, extends from the southwest to the northeast direction in the slope-proximal part of the Davraz hummocky moraine field (Figures 4.4 and 4.9). The presence of a fault zone can boost karstification by creating pathways for groundwater seepage, thus expediting the dissolution process. In the slope-proximal part of the Kırıyayla section (Figure 4.1a), the depression along with the Holocene fault could be linked to accelerated dissolution in this region (Figure 4.9). The dissolution can be attributed to the meltwaters (Bennett & Glasser, 2009; Maire, 1990; Smart, 1983, 2004) from the retreating paleoglacier on the slopes of Mt. Davraz. Cross-sections (Figure 4.9) reveal that the depression along the fault is likely associated with postglacial karstification. We posit that this depression evolved through a process analogous to the formation of solution dolines (cf. Ford & Williams, 2007). Furthermore, based on the topographic features observed, it is inferred that the area in question (southeast part of the Kırıyayla section) served as the discharge point for the meltwater (Figure 4.7). This process might also affect the relief of the adjacent hummocky moraine fields (e.g., Salcher et al., 2010). As a consequence, the depression in this area has become deeper. Currently, seasonal debris flows are still witnessed along this relevant route of flow.

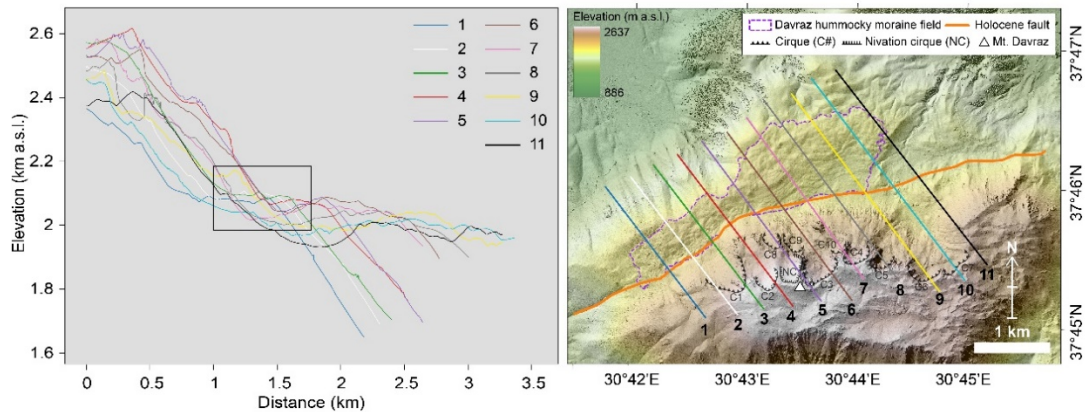


Figure 4.9 : The graph (left) and digital surface model (DSM) (right) display the profiles taken nearly perpendicular to the Holocene fault that traverses between the base of the north slope of Mt. Davraz and the hummocky moraine field. The Holocene fault is indicated by the bold orange line on the DSM.

Sample #8 is indicative of the second phase of the Davraz paleoglacier deglaciation in the Kırıyayla section. It appears that the glacier retreated in two directions in this area - towards the west, in the direction of Sample #2 (17.2 ± 1.8 ka; 2101 m a.s.l.), and

south southeastwards towards the slope. Consequently, this led to bidirectional deglaciation: (i) retreat in the direction of the slope and (ii) down-wastage due to ice disintegration at the central part of the Davraz paleoglacier (Figure 4.8c,d). This bidirectional deglaciation also may have resulted in a high amount of meltwater discharge into the area, accelerating surface lowering through dissolution. Based on the average of samples #3 and 8, we can infer that deglaciation in this area began around 21.8 ± 2.4 ka. Eventually, the paleoglacier patches in the Kırıyayla section presumably disappeared before 17.2 ± 1.8 ka (Sample #2) (Figure 4.8e).

As mentioned before, the Kundurpinarı section represents a different area of cover compared with the Kırıyayla section. Glaciers covered with debris are composed of a core of glacier ice that is covered by a layer of surface debris. The layer of surface debris, which can be several meters thick, typically results from a combination of rockfall onto the glacier ice and/or sublimation of ice-containing debris (Milkovich et al., 2006). Accordingly, given the formation of the piedmont glaciers, they can be covered by thick debris and thus develop high-relief hummocky moraine topographies during the down-wasting of glacier ice. Based on the longitudinal extent and high hummocky moraine relief of the Kundurpinarı section, we are considering that the paleoglacier was thicker here (Figure 4.7). Besides, the dissolution along the Holocene fault seems not intense in the Kundurpinarı section as observed in the Kırıyayla section. Here, the hummocky moraines have quite distinct morphology along with the Holocene fault. These moraines exhibit smaller dimensions and a more compact arrangement compared to other moraines. The distribution of the moraines observed in the area may reasonably be attributed to the deglaciation pattern (i.e., as the paleoglacier retreats towards the slope) of the paleoglacier.

Our findings may indicate that the ages obtained in the Kundurpinarı section are relatively young. We assume that deglaciation occurred in two phases here. Sample #1 (18.7 ± 1.8 ka; 2102 m a.s.l.), as a second paleoglacier patch, is located on probably the first surface exposed in this section (Figure 4.8d). This area underwent disintegration from both the southwest and northeast, as well as from samples #4 and #2, respectively. We believe that the intense glacial deposition in this area is due to its proximity to Cirque #1 (C1), the largest cirque on Mt. Davraz. The morphology of the moraines observed in Sample #2 and its immediate vicinity (Figure 4.7) suggests that this area corresponds to the third patch of the paleoglacier. This patch, which retreated

westward from the Kırıyayla section, also appears to have retreated northeastward from the high-relief area where Sample #1 is situated (Figure 4.8d,e). Finally, the fourth paleoglacier patch which consists of Sample #4 (17.1 ± 2.6 ka; 2083 m a.s.l.), probably melts out synchronously with the third paleoglacier patch (Figure 4.8e). This area also has the clearest boundaries of the hummocky moraine field with two terminal moraine-like loops that include Sample #4. On the other hand, Sample #6, considered an outlier, is situated approximately 250 m northwest of Sample #4 on an adjacent moraine loop. The age of Sample #6 is attributed to its potential detachment from the bedrock through frost-shattering in its original position. Subsequently, it may have rolled onto the glacier, being transported supraglacially to its current location during the flow. Throughout this period, it could have been exposed to cosmic rays, leading to the potential accumulation of cosmogenic ^{36}Cl over tens of thousands of years. Ultimately, we consider that the average initiation timing of deglaciation in the Kundurpınarı section is approximately 17.7 ± 2.2 ka (weighted means of samples #1, 2, and 4), occurring in the early Late-glacial period (Figure 4.8f). In conclusion, based on morphological evidence and the obtained TCND ages, the paleoglacier over the Davraz hummocky moraine field appears to have disappeared approximately from the early Late-glacial period (c. 17 ka). However, some morphostratigraphic evidence suggests the presence of more recent glacial advances in this area. Terminal moraines mapped in front of Cirque 1 and 2 (Figure 4.4) can be considered as evidence for these advances.

4.4.3 Wind-induced snow redistribution effect on glaciation

Temperature and precipitation represent the principal climatic parameters essential for the formation of glaciers. Nevertheless, glacier growth tends to occur on the leeward side of prevailing winds. (Sanders et al., 2013). Snow that is blown by the wind accumulates on the leeward side of crests without any depressions and in sheltered basins (Seppälä, 2004). It is also possible to form glacierets on the leeward slopes by wind-blown snow as well as snow avalanches (Benn & Evans, 2010). These observations highlight the phenomenon where windward slopes are frequently cleared of accumulated snow, and the turbulent conditions on these slopes intensify heat transfer and facilitate ablation processes (Evans, 1977). The aspect of the cirques, which refers to the direction of cirque glacier development, provides valuable insights into the correlation between wind direction and glacier development at this point.

When considering the influence of local wind patterns on Mt. Davraz, it is noteworthy that winds might be affected by the glacier developments on Mt. Davraz (Figure 4.10a).

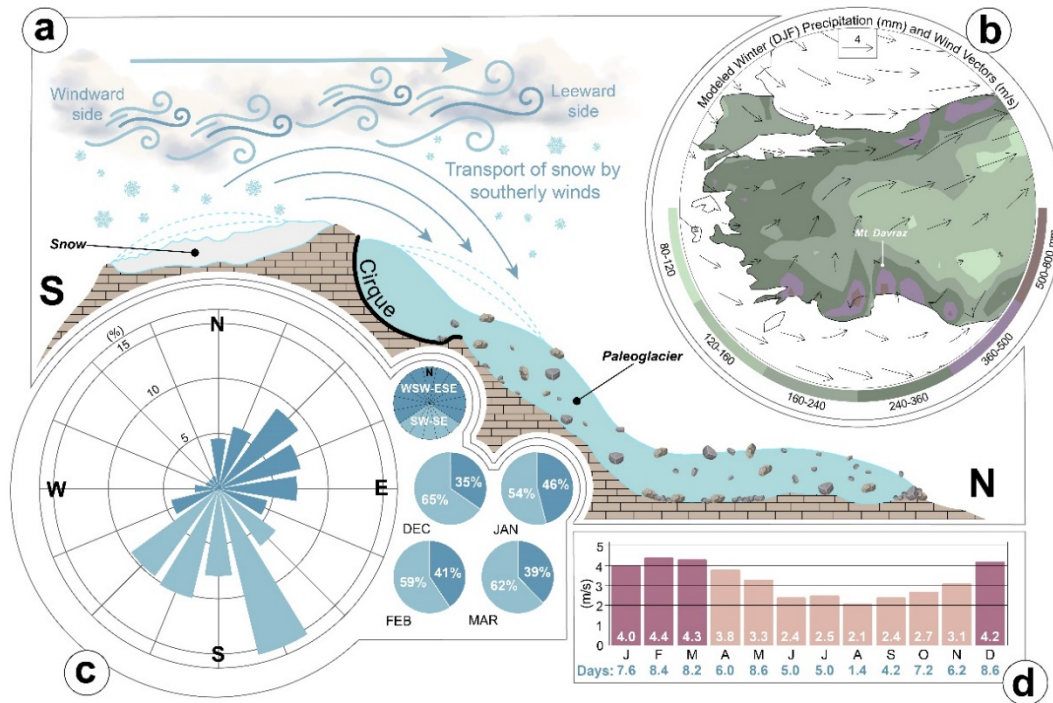


Figure 4.10 : Wind influence on snow accumulation over paleoglaciers of Mt. Davraz: (a) a schematic illustration depicting the influence of wind on the glacial formation of Mt. Davraz; (b) the spatial pattern of simulated precipitation (mm) in Türkiye during winter (DJF) based on a 10-year average (1990–2000). Additionally, displayed are the 10-year seasonal average of 850 hPa wind vectors (m/s) from NCEP/NCAR reanalysis on the left and the control simulation on the right (Bozkurt & Şen, 2011); (c) on the left, a wind rose diagram illustrating the yearly frequency and direction of wind blowing at the Davraz Ski Center weather station. On the right, pie charts indicate the percentage of wind direction frequency during winter (DJF) and March; (d) the columns represent the monthly average wind speed recorded by the Davraz Ski Center weather station. Below the graph, the average number of days with strong winds (10.8–13.8 m/s, Beaufort 6) per month is shown in blue. Notably, the average wind speed and the average number of days with strong winds are presented for the winter (DJF) and March when SW-SE winds are prevalent.

The geomorphology of Mt. Davraz indicates a notable influence of wind patterns on the formation and evolution of glaciers in the area. According to Evans et al. (2021), the presence of westerly winds (Figure 4.10b) and moisture-bearing southerly winds (Figure 4.10c) likely exerted an influence on the snowfall patterns experienced by Mt. Davraz during the Quaternary period. The cirques on Mt. Davraz collectively developed in the NNW-N direction. Consequently, it can be inferred that the

development of these cirques is influenced by local winds predominantly originating from the SSE-S direction.

By examining the current wind directions and frequencies (Figure 4.10c), as well as wind speeds (Figure 4.10d) between December and March on Mt. Davraz, we may gather insights into the role of wind for the Quaternary glaciations. Certainly, to assert such a proposition, one must initially acknowledge the assumption that the paleo-wind parameters are either identical or at least comparable to the present-day conditions. However, based on the relationship between the aspect of the cirques and the current wind directions, it is reasonable to conclude that the current and paleo-wind directions are likely similar. In conclusion, we believe that winds exert a significant influence on the relatively small-scale yet intense paleo-piedmont glaciations on Mt. Davraz.

4.4.4 Glacial records in the nearby region

Mt. Dedegöl (2992 m a.s.l.) and Mt. Barla (2798 m a.s.l.) are two mountains close to Davraz (Figure 4.11). Mt. Dedegöl is located ~50 km east of Mt. Davraz, while Mt. Barla, situated ~30 km north of Mt. Davraz, stands as the northernmost peak in this Western Taurus mountains. Both mountains exhibit cirque-valley-type glaciation, which has played a significant role in shaping their overall geomorphology. However, significant differences can be observed between these two mountains. Mt. Dedegöl stands out with its considerably higher relief presenting an intense glaciation compared to Mt. Barla. On the other hand, the glacial geomorphology of Mt. Barla appears to be more marginal (Altınay et al., 2022; Ardos, 1977a). However, when compared to Mt. Davraz, these differences are even more pronounced. Topographic factors on Mt. Davraz have hindered the development of glacial valleys. Instead, the glaciers that formed within the cirques accumulated and spread at the mountainside, giving rise to a piedmont-type glaciation. Therefore, Mt. Davraz exhibited a predominance of glacial deposition characterized by hummocky moraines.

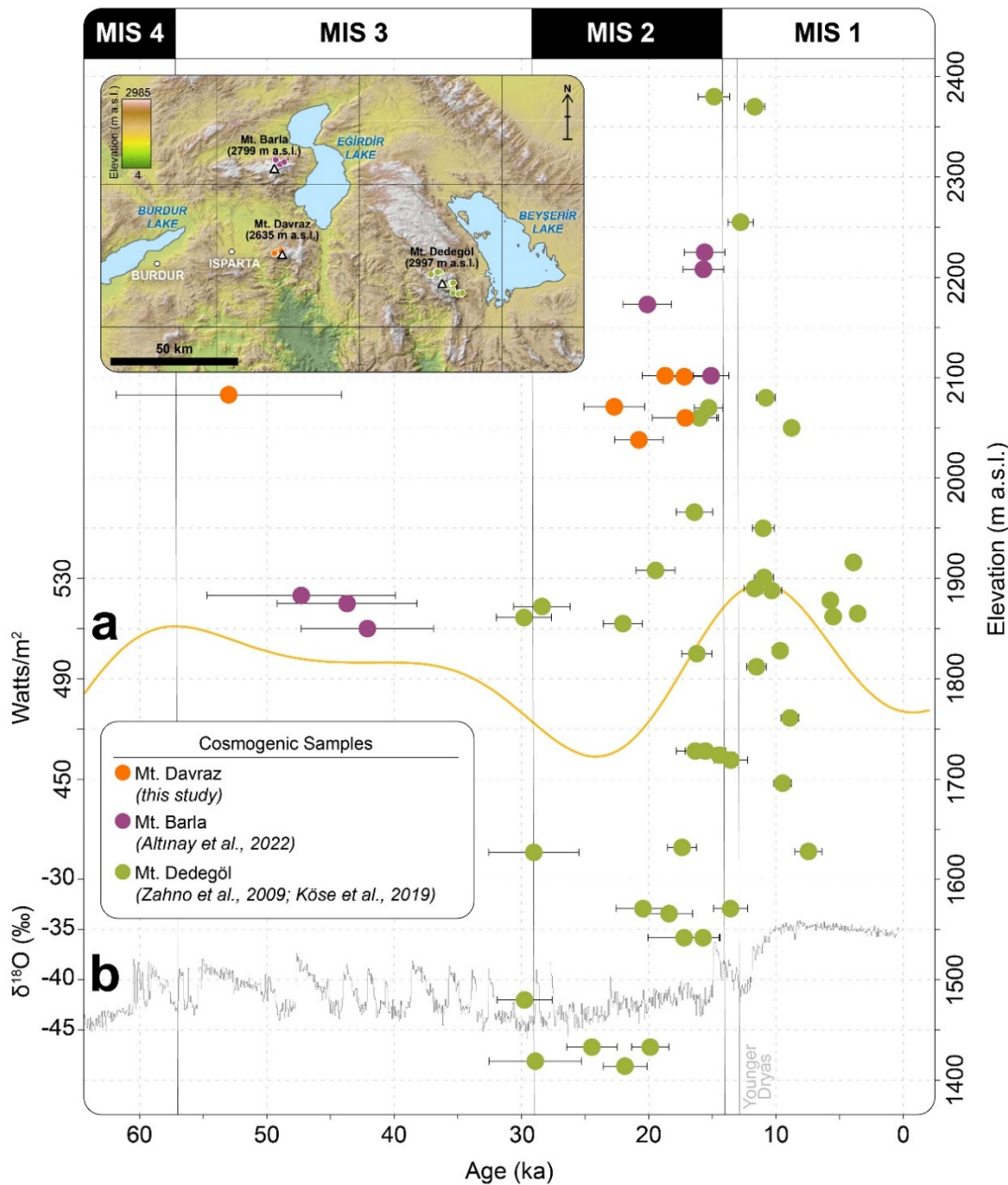


Figure 4.11 : The cosmogenic ages of Mt. Davraz are shown together with the cosmogenic ages of the nearby Dedegöl and Barla mountains. (a) the mean daily insolation at 65° N on the summer solstice data from Laskar et al. (2011) and (b) the oxygen isotopic profile from the NGRIP ice core (North Greenland Ice Core Project members, 2004). The map at the top left displays the locations of Mt. Davraz, Mt. Dedegöl, and Mt. Barla with their cosmogenic age locations.

Köse et al. (2019) reported that the primary source of precipitation in the current area on Mt. Dedegöl is the prevailing easterly winds. Additionally, they noted that these easterly winds played a significant role in the formation of cirque and valley glaciers during glacial periods. It can be regarded as one of the most heavily glaciated mountains in the Western Taurus region. The first dating study conducted in this area

was by Zahno et al. (2009). They found evidence that the glaciers formed in three distinct periods in the W-E-oriented Muslu Valley: during the global LGM (24.3 ± 1.8 ka), late LGM (21.5 ± 1.5 ka), and in the Late-glacial period (15.2 ± 1.1 ka). This study also holds a place in the literature as the first and only cosmogenic ^{10}Be and ^{26}Al isotopes surface exposure dating in the Taurus Mountains. Köse et al. (2019) obtained similar age results for the Sayacak and Karagöl valleys using cosmogenic ^{36}Cl . According to their findings, there are correlated age of deglaciations with Mt. Davraz from the late LGM period (19.7 ± 1.6 ka) and the Late-glacial period (16.4 ± 0.7 ka and 12.0 ± 1.0 ka).

In contrast to Mt. Dedegöl, precipitation in Mt. Barla (and Mt. Davraz) is primarily influenced by west winds (Evans et al., 2021). However, the glaciations of Mt. Barla overlap only within the Late-glacial boundaries with Mt. Davraz. The only TCND age shows that final deglaciation occurred during Late-glacial (16.6 ± 2.6 ka) (Altınay et al., 2022). Therefore, it can be concluded that final deglaciations on Mt. Barla, as well as on Mt. Davraz, were predominantly restricted to the Late-glacial (c. 19 to 11.7 ka), in contrast to Mt. Dedegöl, where glacial activity extended to younger periods.

Despite the variations, the glacial histories of these three mountains largely overlap (Figure 4.11). For instance, summer insolation (Figure 4.11a) between ~25 and 13 ka reveals a potential commonality among these three mountains. In this specific range, which fluctuates roughly from 460 to 530 w/m^2 , especially above 1800 m a.s.l., a monotonic relationship appears to exist between the insolation line and glacier retreat concerning the elevation of cosmogenic samples. However, upon examining the NGRIP Ice core data (Figure 4.11b), the monotonic relationship terminates just before the Younger Dryas (at the onset of the Bølling-Allerød interstadial; 14.6–12.9 ka, Rasmussen et al., 2014) based on the grouped ages. From this juncture onward, according to the existing chronology, neither Mt. Davraz nor Mt. Barla can be associated with any glacial retreat.

Based on previous studies, the LGM ELAs of Mt. Dedegöl (Çilgin, 2012), Mt. Barla (Evans et al., 2021), and Mt. Davraz (Evans et al., 2021) are 2230 m a.s.l., 2450 m a.s.l. and 2233 m a.s.l., respectively. Cosmogenic sample elevations on these three mountains range from 1400 to 2400 m a.s.l. In contrast to Mt. Dedegöl, the samples from Barla and Davraz mountains, representing the LGM and Late-glacial period, are concentrated at elevations of roughly between 2000 and 2250 m a.s.l. The wide

elevation range of the LGM-aged samples on Mt. Dedegöl suggests a more extensive glaciation in this region (Figure 4.11). In contrast, the mean elevation of the Late-glacial samples on Mt. Barla (Altınay et al., 2022) is 273 m below the LGM ELA, while the mean elevation of the samples on Mt. Davraz is 168 m below (96 m below the LGM ELA calculated with AABR method) the LGM ELA of the mountain. Therefore, it can be concluded that there was marginal glaciation in these two mountains.

The key advantage of Mt. Davraz among these three mountains is its positioning along the path of humid air masses originating from the west and south (Figure 4.10b). In contrast to Evans et al. (2021), we hold the view that the impact of moisture-bearing southerly winds on Mt. Barla is rather limited due to the location of Mt. Davraz. We believe that Mt. Davraz receives the majority of the precipitation by intercepting the air masses arriving from the south towards Mt. Barla. Accordingly, we hold the belief that the primary factor contributing to the formation of glaciers on Mt. Barla is the influence of westerly winds. It is known that the paleoglaciers on Mt. Dedegöl developed under the control of easterly winds (Evans et al., 2021; Köse et al., 2019). In this regard, Mt. Davraz holds a more advantageous position for glacier development, despite its relatively smaller scale, as it benefits from the convergence of humid air masses originating from both the south and west.

4.4.5 Piedmont glaciations and hummocky moraines in the Mediterranean region

Piedmont glaciations are infrequent in the Mediterranean region during Quaternary glaciations (e.g., Çiner et al., 2015; Fernandes et al., 2017). Additionally, while hummocky moraines are observable on valley floors, the chances of encountering extensive areas of hummocky moraines, particularly those originating from ice sheets, are low.

Prior to this investigation, Geyikdağ stood as the singular mountain in Türkiye known for hosting piedmont glaciers during Quaternary glaciations. It is situated 165 km northwest of Mt. Davraz. The paleo-piedmont glaciers here ($\sim 25 \text{ km}^2$) were five times larger than those on Mt. Davraz ($\sim 4 \text{ km}^2$). Smaller glaciers, known for their higher sensitivity (Bahr et al., 1998; Huss & Fischer, 2016; Oerlemans & Fortuin, 1992; Zekollari et al., 2020), suggest a rapid glacial retreat on Mt. Davraz. We propose that

the Davraz hummocky moraine field has likely been ice-free after the early Late-glacial (17.7 ± 2.2 ka).

Çiner et al. (2015) identified a Late Pleistocene ice cap on Geyikdağ, attributing piedmont glaciers and hummocky moraine formation to it. In contrast, our study suggests that the piedmont glaciers on Mt. Davraz originated from cirque glaciers, given the absence of glacial evidence on the summit plateau. Satellite images resembling a hummocky moraine region contradicted our field observations, revealing a doline-uvala complex formed by karstification. Insufficient snow accumulation due to the windward position is believed to hinder glacier formation in this area (Figure 4.10).

A thorough investigation conducted by Smith et al. (1997) on Mt. Olympus (2917 m a.s.l.) in Greece revealed that the initial glacial phase was characterized by piedmont lobes advancing toward lower elevations. The age of this glaciation, dating back to 200 ka, was determined through soil stratigraphy correlation. Subsequent glaciations were confined to valley glaciations. However, Woodward et al. (2004) argue that the chronology in this study is not reliable and lacks internal consistency, thereby limiting its value as a source of paleoclimatic information. On the other hand, relying on the ages derived from two moraine boulders (12.0 ± 0.9 ka and 12.1 ± 0.8 ka) in the hummocky moraine area within the Megala Kazania cirque on Mt. Olympus, Styllas et al. (2018) believe that these ages signify the conclusion of the second Late-glacial phase (~ 12.5 ka) of glacial activity on Mt. Olympus. They also identified hummocky moraine areas in the Pindus Mountains in Greece (Hughes et al., 2006). The U-series dating results obtained from cement formed in tills around hummocky moraines indicate that the minimum ages of the tills fall between MIS 5a interstadial and MIS 6. Nevertheless, the formation of hummocky moraines in Greece is not related to either piedmont glaciations or ice sheets, as seen in Mt. Davraz.

Examples of (paleo-) piedmont glaciations and moraine formation are also documented in the Dinaric Mountains, situated in the central part of the Mediterranean region along the Adriatic Sea. According to Çiner et al. (2019), on Mt. Čvrsnica (2226 m a.s.l.), piedmont glaciers advanced towards Blindje Polje, where they created hummocky, lateral, and terminal moraines. Ages derived from five hummocky moraine boulders from the Svinjača area indicate that the glacial masses reached their maximum of approximately 22.7 ± 3.8 ka. This conclusion was drawn using the age

of the oldest among the five moraine boulders. They also asserted that this result is spatiotemporally consistent with the lateral (13.2 ± 1.8 ka) and terminal (13.5 ± 1.8 ka) moraine ages within the polje. The onset of deglaciation on the Svinjača hummocky moraine area correlates with the deglaciation of the Davraz paleoglacier.

4.5 Conclusion

The identification of paleo-piedmont glaciers, which give rise to the hummocky moraine field on the northern slope of the mountain, represents a significant contribution to the literature, particularly in the realm of Mediterranean glaciations. Such features are relatively uncommon in mid-latitudes, making their discovery on Mt. Davraz all the more noteworthy.

Based on the ages derived from six cosmogenic samples taken from the Davraz hummocky moraine field, it was determined that the region experienced two distinct phases of glacial retreat. The first retreat of the Davraz paleoglacier commenced approximately 21.8 ± 2.4 ka (late LGM) years ago in the eastern part (Kıryayla section). Subsequently, the western (Kundurpinarı section) part of the paleoglacier retreat began approximately 17.7 ± 2.2 ka (early Lateglacial) years ago. Nevertheless, the terminal moraines in front of some cirques indicate morphostratigraphically younger glacial advances.

Despite its relatively small scale, topographical effects around Mt. Davraz are considered to be decisive factors in creating favorable conditions for the development of piedmont glaciers. During this period, the influence of moist air masses originating from western and southern winds played a significant role in the glaciation of Mt. Davraz. Another important impact of the southerly winds was their ability to transport snow from the windward side to the leeward side of the slopes. We believe these winds directly contributed to the formation and development of cirque glaciers on the northern slopes.

Mt. Barla and Mt. Dedegöl, as nearby mountains, do not share any similarities with Mt. Davraz in terms of glacial geomorphology. Nevertheless, their glacial chronologies are correlated during the global LGM (26.5 to 19 ka) and Late-glacial (c. 19 to 11.7 ka). Mt. Davraz exhibits significant geomorphological and geochronological similarities to Geyikdağ in Türkiye. However, these resemblances

do not extend to piedmont glaciations and extensive hummocky moraine fields in the Mediterranean region. The remaining piedmont glaciations in the region are either old (e.g., pre-LGM) or have not demonstrated the potential to form extensive moraine areas. This is likely attributed to the topographies of these areas which do not support the formation of an ice sheet by piedmont glaciers, as seen in Mt. Davraz.

Overall, despite the challenge of dating hummocky moraines with carbonate lithology, our limited data evidence a consistent chronology both with global and local LGM. With a limited number of samples, general conclusions about the deglaciation of the study site could not be made exclusively. Future studies in the region may focus particularly on understanding karst processes of carbonate rocks on glaciated regions with a larger set of samples. However, we believe that the established chronology also enlightens and emphasizes how much the topography and climate can be effective on glaciations, even in small-scaled mountains. The rarity of the relationship between piedmont glaciations and hummocky moraine fields in a mid-latitude region like the Mediterranean serves as evidence for this assertion.



5. TESTING SCHMIDT HAMMER EXPOSURE-AGE DATING (SHD) ACCURACY: AN EXPERIMENTAL STUDY ON DIFFERENT LITHOLOGIES IN WESTERN TAURUS

5.1 Introduction

The Schmidt hammer (SH), originally developed for engineering purposes by Ernst Schmidt to measure concrete hardness and deterioration (Schmidt, 1950, 1951), has been used to measure the hardness of rocks/concrete in natural/anthropogenic environments for over half a century (Yaalon & Singer, 1974; Day & Goudie, 1977). A rebound value (R-value), measured on a scale of 1 to 100, serves as an artificial numerical index of the Schmidt hammer. Mechanical engineers more accurately identify R-value as ‘compressive strength’, but in earth-science contexts, it is often simplistically labeled as ‘surface hardness’ (Matthews & Winkler, 2022). In chronological applications, R-values are utilized to gauge time-dependent alterations in the extent of weathering, presupposing that as rocks undergo weathering, chemical alterations and the development of near-surface microfractures amalgamate to diminish surface hardness and compressive strength (McCarroll, 1989). Thus, it is anticipated that there will be an inverse relationship between the R-value and the age of the exposed surface. In other words, surfaces with higher R-values are expected to be exposed to atmospheric conditions earlier than those with lower R-values.

The first Schmidt hammer (relative-age) dating study within the scope of geosciences was conducted by Matthews & Shakesby (1984). Especially since the 21st century, the popularity of the Schmidt hammer dating has been steadily increasing. These studies have been conducted particularly on glaciated areas, as well as on surfaces where various geological and geomorphological processes are effective like moraines (Aa & Sjøstad, 2000; Winkler et al., 2003; Matthews & Winkler, 2011), glacially polished bedrocks (Matthews & Owen, 2010; Böhlert et al., 2011; Dąbski, 2014), rock glaciers (Rode & Kellerer-Pirklbauer, 2012; Kłapyta, 2013; Scapozza et al., 2021), rock avalanches (Clark & Wilson, 2004; Aa et al., 2007; Longhi et al., 2024), pronival ramparts (Matthews & Wilson, 2015; Matthews et al., 2017; Marr et al., 2019), alluvial

and colluvial fans (Decaulne et al., 2007; McEwen et al., 2020, Matthews et al., 2020) and other various landforms.

The use of the Schmidt hammer for dating purposes is approached in two fundamentally different ways by Matthews & Winkler (2022). In the Schmidt hammer relative-age dating (SRD) method (1), there is no calculation of a numerical age calibration curve, whereas in the Schmidt hammer exposure-age dating (SHD; another commonly used abbreviation being SHED, e.g, Tomkins et al., 2016; Dortch et al., 2016; De Marcos et al., 2019) method (2), a calibration curve is developed for dating purposes or an existing age calibration curve is utilized. The idea of using the Schmidt hammer for numerical-age dating, which involves estimating the age in years on a ratio or interval time scale (cf. Colman et al., 1987), was initially proposed at least 20 years ago (Betts & Latta, 2000; Haeberli et al., 2002; Shakesby et al., 2006). Since then, SHD has been utilized alongside various numerical dating methodologies including terrestrial cosmogenic nuclides (TCNs), lichenometry, radiocarbon (^{14}C), optically stimulated luminescence (OSL), and thermoluminescence (TL). In essence, SHD is a calibrated-age dating method relying on the extent of weathering observed on rock surfaces.

In the realm of civil engineering, Schmidt hammer readings (R-values) indicating potential concrete defects prompt the collection of samples for thorough analysis, while in geomorphological studies, a significant quantity of R-values is utilized (McCarroll, 1989). A large quantity of R-values is important in SHD because even minor variations in mean R-values can lead to statistically significant results. There is no specific number of impacts for R-value measurements per surface (e.g., bedrock or boulder surfaces); however, most studies have shown researchers typically make readings between 1 and 30. Essentially, there appears to be a roughly inverse relationship between the number of impacts and the surfaces being measured. As per Matthews & Winkler (2022), in scenarios where boulders are abundant and surfaces exhibit mixed ages, the preference may lean towards favoring one impact from each of a relatively large number of boulders (Matthews et al., 2014, 2015). Conversely, when boulders are scarce or large, it may be more appropriate to use a larger number of impacts per boulder (Matthews et al., 2013; Nesje et al., 2021).

In this section of the thesis, the results obtained from the application of the SHD for the first time in Türkiye, along with the encountered challenges and prospective

research, are discussed. The study was conducted in the glaciated areas of Mt. Sandıras, Karadağ, and Akdağ, located within the Western Taurus Mountains in southwestern Türkiye. Surface exposure ages of undated surfaces exposed to glaciation on these mountains were attempted to be calculated by evaluating Schmidt hammer measurements (R-values) and TCND ages together.

5.2 Physical setting of the study area

The mountains of the study area, from west to east, are Mt. Sandıras ($37^{\circ} 4' 52''$ N, $28^{\circ} 50' 17''$ E, 2295 m a.s.l.), Karadağ ($36^{\circ} 52' 10''$ N, $29^{\circ} 18' 37''$ E, 2418 m a.s.l.), and Akdağ ($36^{\circ} 32' 38''$ N, $29^{\circ} 34' 5''$ E, 3014 m a.s.l.) are positioned in a northwest to southeast alignment, delineating the western extremity of the Mediterranean Taurus (Figure 5.1). The Dalaman River flows into the Mediterranean Sea between Mt. Sandıras and Karadağ, while the Eşen River flows into it between Karadağ and Akdağ. The distances between the summits of the mountains and the coastline are roughly 30 km. Accordingly, it can be said that the elevation changes along the coastal areas are quite abrupt, much like in the rest of the Western Taurus Mountains.

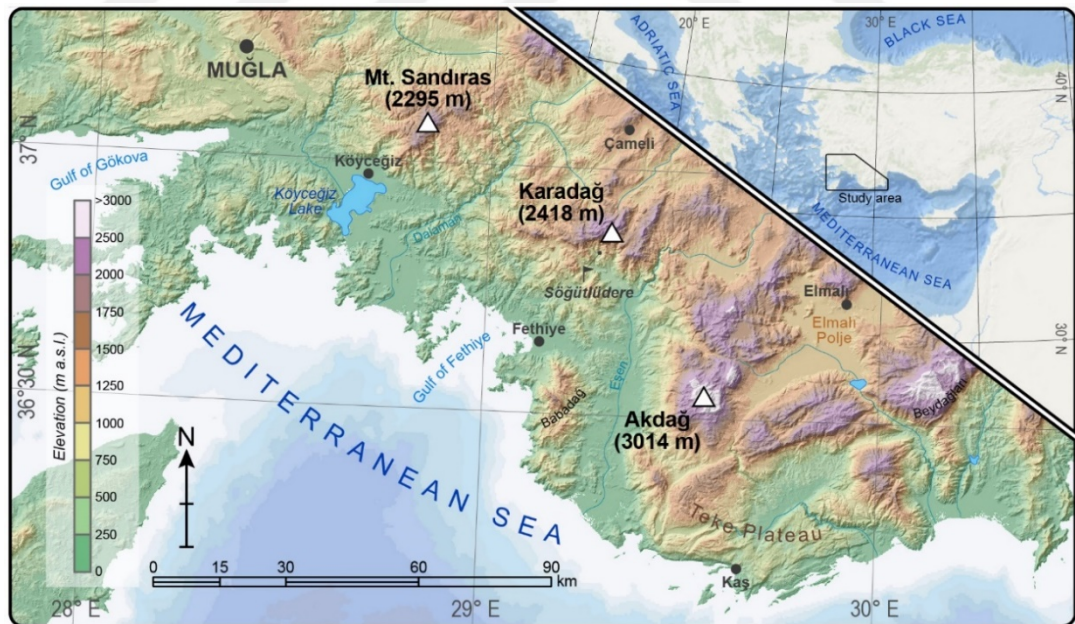


Figure 5.1 : Location map showing the study sites. Except for Söğütlüdere weather station, the stations in Köyceğiz, Fethiye, Çameli, Elmalı, and Kaş are located in the city centers.

The Western Taurus Mountains are characterized by Mesozoic autochthonous carbonate platforms that have been overthrust by the Lycian Nappes, dating from the

Late Cretaceous to the Pliocene period (Şenel, 1989). The carbonate rocks present in the region, particularly the insoluble units within the complex structural belt of the Lycian nappes, have played a significant role in the geomorphological development. The Taurus Mountains, generally composed of carbonate rocks, exhibit contrasting lithologies in some locations. Mt. Sandıras is one of these instances. The geological formation visible on the mountain is the upper section of the Lycian Allochthons, referred to as the Lycian Peridotite Thrust Sheet (Collins & Robertson, 1998). It primarily comprises serpentized harzburgite, with minor occurrences of pyroxenite, pediform dunite, and chromitite (Kaaden, 1959; de Graciansky, 1967; Engin & Hirst, 1970; Collins, 1997). Karadağ and Akdağ, on the other hand, exhibit the general lithological characteristics of the Western Taurus Mountains: while the areas near the glaciated peaks are predominantly composed of carbonate rocks, the lower parts contain impermeable units associated with the Lycian allochthons (Şenel, 1997). In Karadağ, the most extensive units consist of Middle Triassic dolomites and Jurassic-Cretaceous limestones. Similarly, Akdağ is mainly composed of Jurassic-Cretaceous Lycian allochthons (Koşun et al., 2009) containing pelagic and neritic limestones (Colin, 1962; İslamoğlu & Taner, 2002). Glacial geomorphology has also significantly developed on these lithologies in Karadağ and Akdağ.

The Mediterranean Taurus, which includes the Western and Central Taurus mountains, is a transitional zone that combines elements of the continental and Mediterranean climate characteristics. Because of the impact of the Mediterranean climate, temperatures along the south-facing (coastal) slopes stay higher year-round than along the north-facing (inland) slopes. Although the entire region of the Mediterranean Taurus, precipitation patterns are mostly determined by orography. SW-NE-oriented Mediterranean cyclones, which play an important role in the occurrence of orographic rainfall, are most active in the region during the winter months (Almazroui et al., 2015). The Western Taurus Mountains are the most affected by these cyclones due to their ideal geographical location and morphology. Moreover, extreme orographic precipitation events are more common in the Western Taurus because of the parallel alignment of mountains with the coasts of the Marmaris and Antalya Gulfs (Türkeş et al., 2016).

Mt. Sandıras experiences heavy precipitation due to its location near frontal systems moving northeastward from the Mediterranean (Doğan ve Köse, 2015). According to

the data from the Köyceğiz meteorological station (24 m a.s.l.; Figure 5.1), located southwest of the mountain and near the coast, the mean annual temperature is 18.4 °C, while the mean annual precipitation is 1110 mm (Balcıoğlu, 2021). Bayrakdar et al. (2017) calculated the precipitation data for Karadağ using the data from two meteorological stations (Çameli, 1325 m a.s.l.; Fethiye, 3 m a.s.l.; Figure 5.1) near the mountain and applying the Schreiber formula (Dönmez, 1990). According to their findings, depending on the elevation difference, the annual total precipitation for Karadağ could vary between 1300 and 2000 mm. However, in a recent study by Sarış et al. (2021), the annual precipitation on Karadağ is reported as 1695.6 mm based on one year of observation (15 July 2018 – 15 July 2019). The data were collected by an automatic weather station situated in Söğütlüdere (780 m a.s.l.) on the lower-southern slopes of Karadağ (Figure 5.1). For Akdağ, Bayrakdar (2012) calculated the mean annual temperature and mean annual precipitation based on the data from meteorological stations located in the vicinity of the mountain, including Fethiye (3 m a.s.l.), Elmalı (1095 m a.s.l.), and Kaş (153 m a.s.l.) stations (Figure 5.1). In this respect, the mean annual temperatures for elevations above 2500 m a.s.l. in Akdağ were determined to be 4 °C, while the annual average precipitation for elevations above 2000 m a.s.l. was calculated as 1200 mm.

Mt. Sandıras consists of three north-facing glacial valleys. The largest glacier on Mt. Sandıras extended up to 1.5 km in Kartal Lake Valley (Figure 5.2a) from its cirque and left terminal moraines at an elevation of 1900 m a.s.l. (Sarıkaya et al., 2008). In Karadağ, cirques predominantly oriented towards the north (Figure 5.2b), along with north-directed glacial valleys, serve as evidence of Quaternary glaciations in the region. Similar to the situation on Mt. Sandıras, it is possible to mention paleoglaciers that extended up to 1800 m a.s.l. (Bayrakdar et al., 2017). Compared to Mt. Sandıras and Karadağ, Akdağ exhibits much more well-developed glacial geomorphology (Figure 5.2c). Within the area, seven glacial valleys and 16 cirques have been identified (Bayrakdar, 2012). Based on TCND ages, the last glacial retreat on Mt. Sandıras began around 14.1 ± 2.3 ka (Sarıkaya et al., 2008). The last glacial retreat on Akdağ has been dated to ~17 ka based on similar ages obtained from four different glacial valleys (Sarıkaya et al., 2014).

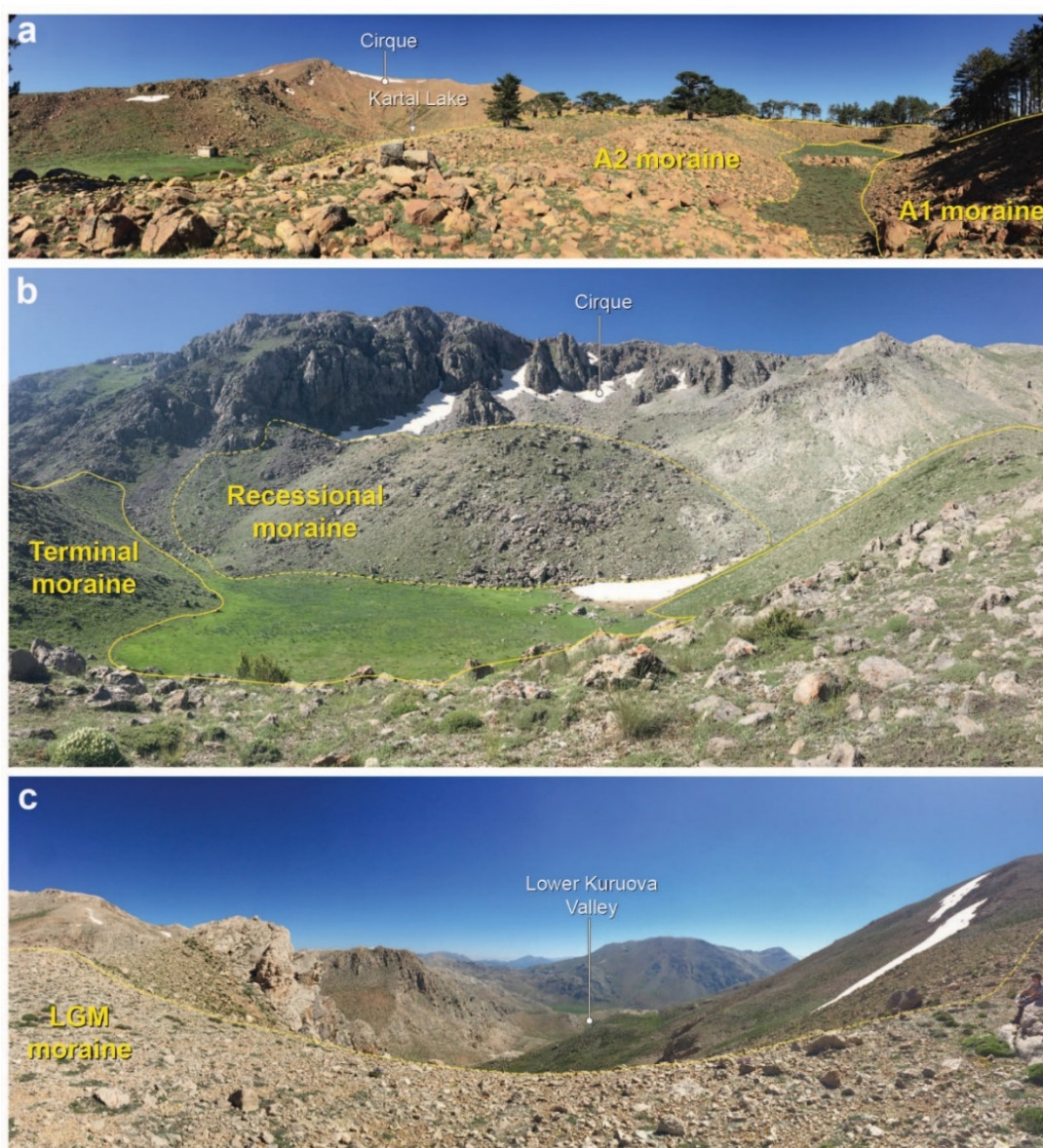


Figure 5.2 : Selected glacial geomorphologies from the study areas: (a) Kartal Lake Valley of Mt. Sandıras viewed from the NE; (b) the top of the terminal moraines of Akdağ Cirque in Karadağ; and (c) Kuruova Valley in Akdağ, viewed along the SW-NE direction from the top of the LGM moraines of Upper Kuruova Valley.

5.3 Schmidt hammer exposure-age dating (SHD) methodology

5.3.1 Sampling strategy

The main sampling strategy involved measuring R-values of each surface with a TCND age (hereafter referred to as "control surfaces") against each surface without an age (hereafter referred to as "test surfaces"). In other words, the R-values of all control surfaces were measured; then, the R-values of test surfaces located a few meters away from each control surface were also measured (Appendix Figure A.1). All control

surfaces (denoted with “-SH”) correspond to a single test surface (denoted with “-X”), except for sample SA22-617 from Mt. Sandıras, which has two test surfaces (i.e., there are SA22-617-X1 and SA22-617-X2). The primary reason for choosing this strategy is to prevent specific sources of error, particularly those related to lithology, by cross-checking R-values for each control-test surface group (Figure 5.3). For instance, carbonate rocks have an unfavorable reputation in Schmidt hammer applications due to their susceptibility to chemical weathering when exposed to atmospheric conditions (Arman et al., 2013; Nasri et al., 2019).

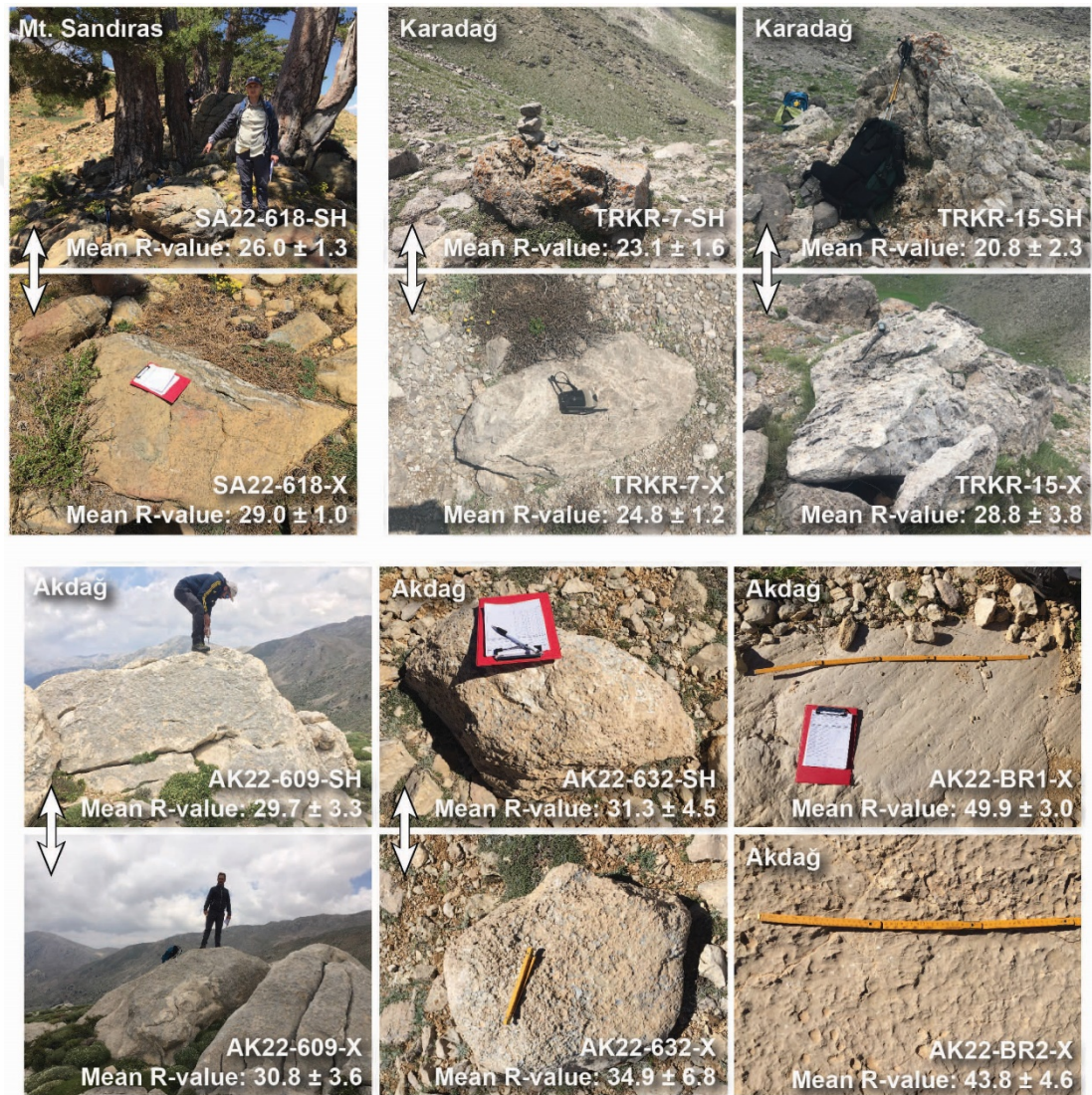


Figure 5.3 : Certain moraine blocks from Mt. Sandıras, Karadağ, and Akdağ with their mean R-values. “BR#” stands for “bedrock” and represents undated glacial bedrock surfaces. Sample IDs ending with “-SH” represent control boulders (i.e., control points), while those ending with “-X” denote test boulders. Double-headed white arrows indicate the control-test boulder sample groups.

Each control surface corresponds to a "control point" (CP) used to generate calibration curves, which will be employed to calculate the SHD ages of undated boulders. The cosmogenic ages used in this study were obtained from ICE-D (Informal Cosmogenic-nuclide Exposure-age Database) Alpine (<https://www.ice-d.org/alpine/>). Certain TCND boulder ages are calculated from the weighted averages of two TCND ages obtained from samples (e.g., samples SA22-616, SA22-617, and SA22-618; cf. Sarikaya et al., 2008). A Proceq N-type Schmidt hammer (Appendix Figure A.2) was used to measure the R-values of sampling surfaces. The measurements were conducted on moraine boulders (n=137; 68 control boulders and 69 test boulders) and glacially abraded bedrock test surfaces (n=2) (Table 5.1). Before and after the fieldwork, 20 impacts were performed with the SH on the marble boulder in front of the Istanbul Technical University Faculty of Mines (41° 6' 18" N, 29° 1' 34" E; Appendix Figure A.3) which is considered the calibration boulder. The R-values were measured as 60.5 ± 0.5 before the fieldwork and 50.5 ± 0.5 after the fieldwork (Appendix Table A.1). During the fieldwork for this study, the R-values of different surfaces exposed outside of glacial processes were also measured using the same SH. Therefore, in addition to those in this study, 3220 additional R-values were measured, bringing the total number of R-values to 6000.

Table 5.1 : The sites in the study areas with conducted SH measurements (*Due to the inability to determine the location of the three control surfaces on the moraine, a total of 120 SH impacts could not be conducted [comprising a total of 6 boulders, including the test surfaces]).

Location	Site	Landform	# of sample surface	# number of total SH impact
Mt. Sandıras	Kartal Lake Valley	A1 moraine	6	120
		A2 moraine	6	120
	Northwest Valley	B1 moraine	2	40
		B2 moraine	5	100
Karadağ	Akdağ Cirque	Terminal moraine	20	400
		Recessional moraine	4	80*
	Karadere Valley	Left lateral moraine	6	120
	Çağılıpınaryayla Cirque	Terminal moraine	6	120
Akdağ	Taşkuzluklu Valley	End moraine	10	200
		Left lateral moraine	6	120
		Right lateral moraine	6	120
		Terminal moraine	14	280
	Karadere Valley	End moraine	2	40
		Right lateral moraine	4	80
		Terminal moraine	16	320
		Lateral moraine	2	40
	Lower Kuruova Valley	Left lateral moraine	4	80
		Right lateral moraine	8	160
		Terminal moraine	8	160
		Lateral moraine	2	40
	Upper Kuruova Valley	Terminal moraine	8	160
		Lateral moraine	2	40
	Central Kuruova Valley	Glacially abraded bedrock	2	40
TOTAL			139	2780

While the control-test boulder group samples are located on Mt. Sandıras (Figure 5.4), Karadağ (Figure 5.5), and Akdağ (Figures 5.6 and 5.7), the glacially abraded bedrock surfaces are located only in Akdağ.

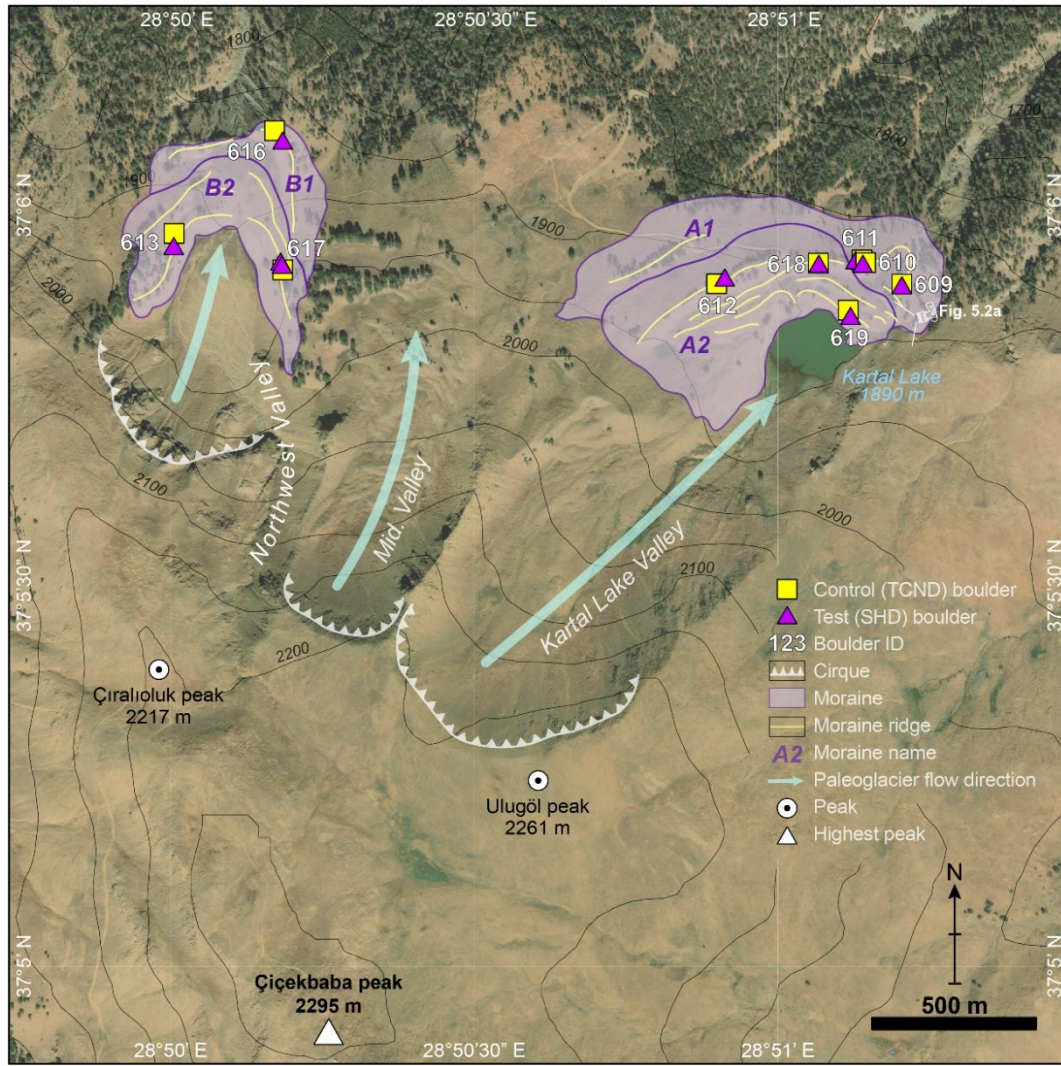


Figure 5.4 : The location of control and test boulders of Northwest and Kartal Lake valleys in Mt. Sandiras. For clarity, only the numerical part of the Sample IDs is provided. The full Sample IDs are formatted as SA22-###-SH for control boulders and SA22-###-X for test boulders. Moraine boundaries and names (A1, A2, B1, and B3) were adapted from Sarıkaya et al. (2008).

Twenty impacts (sample size = 20) were performed per control and test surface. Schmidt hammer impacts were conducted perpendicular to the flat and gently dipping top surfaces of moraine boulders and bedrocks. The average size measurements of the sampled moraine boulders are 182 x 127 x 77 cm (L x W x H).

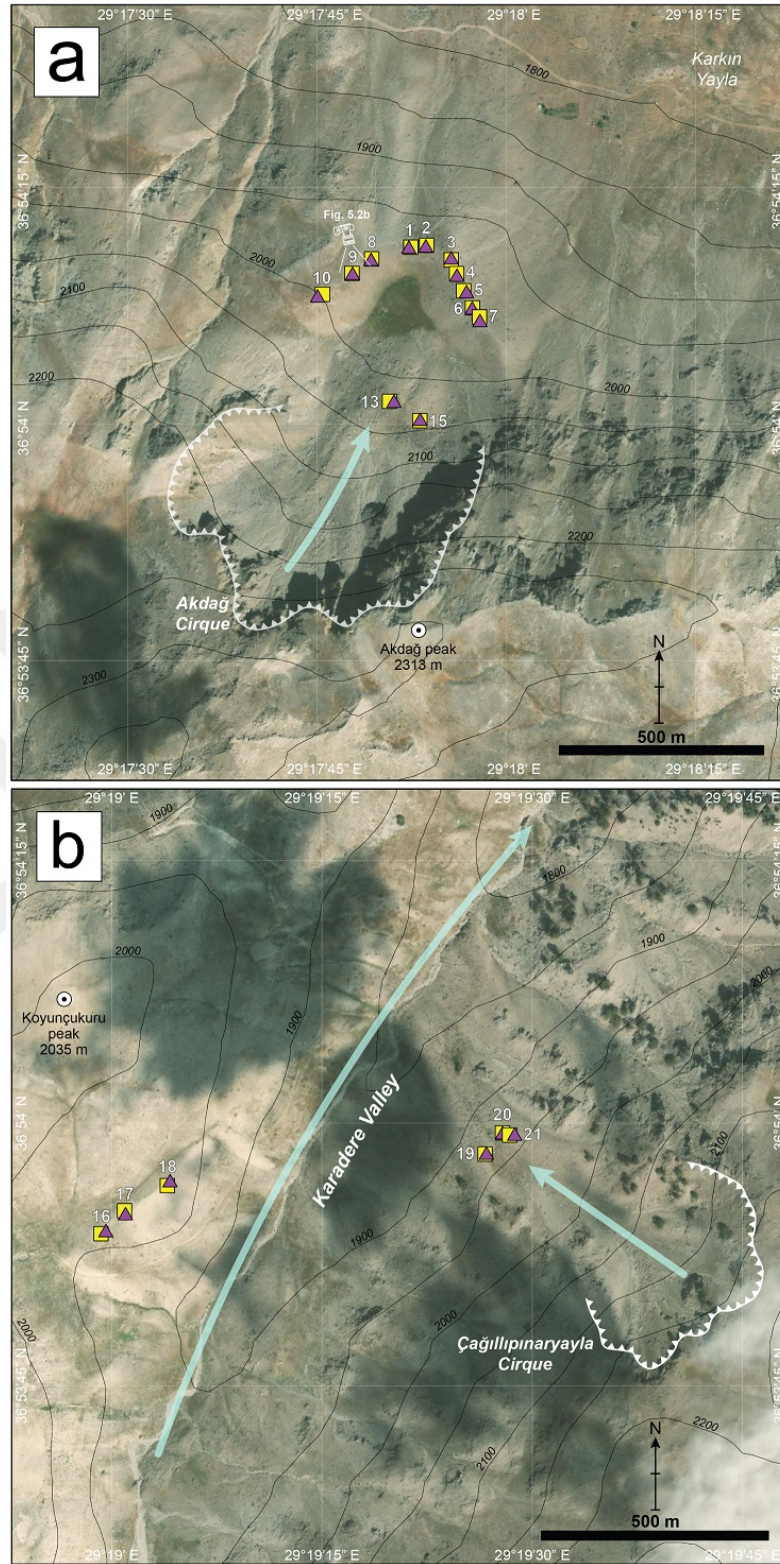


Figure 5.5 : The location of control and test boulders of Akdağ Cirque (a), Karadere Valley, and Çağıllıpinaryayla Cirque (b) in Karadağ. For clarity, only the numerical part of the Sample IDs is provided. The full Sample IDs are formatted as TRKR-###-SH for control boulders and TRKR-###-X for test boulders. Refer to Figure 5.4 for the legend.

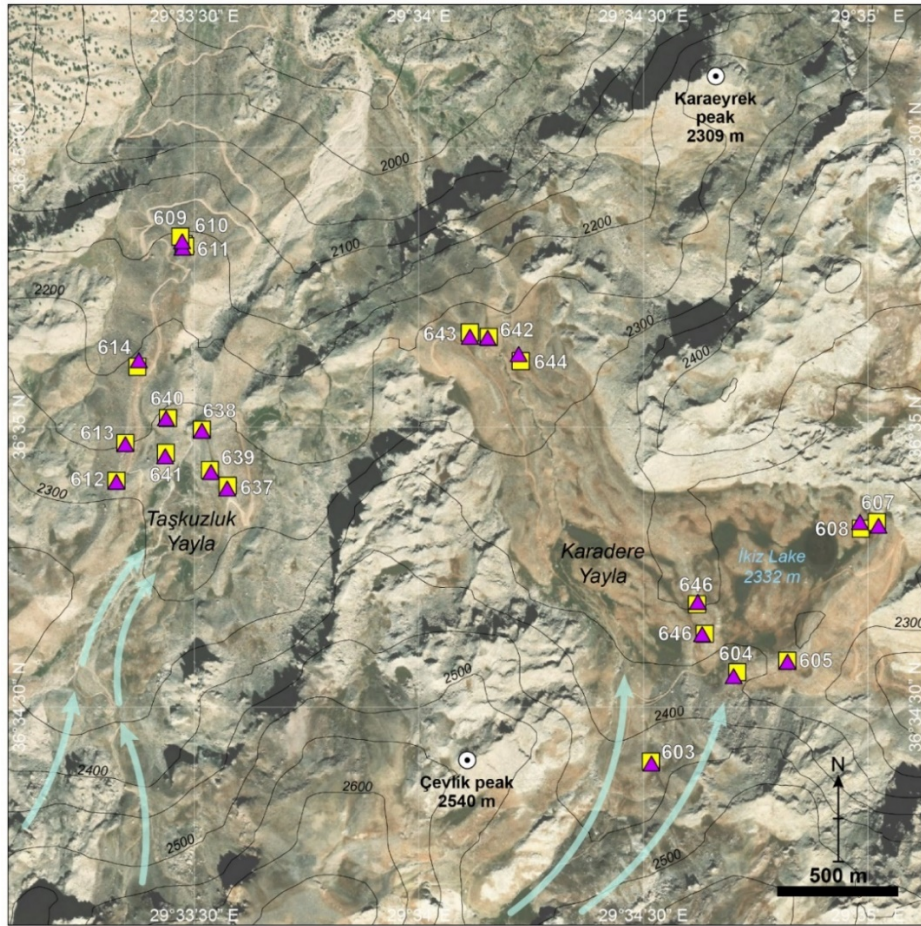


Figure 5.6 : The location of control and test boulders of Taşkuzluklu and Karadere valleys in Akdağ. For clarity, only the numerical part of the Sample IDs is provided. The full Sample IDs are formatted as AK22-####-SH for control boulders and AK22-####-X for test boulders. Refer to Figure 5.4 for the legend.

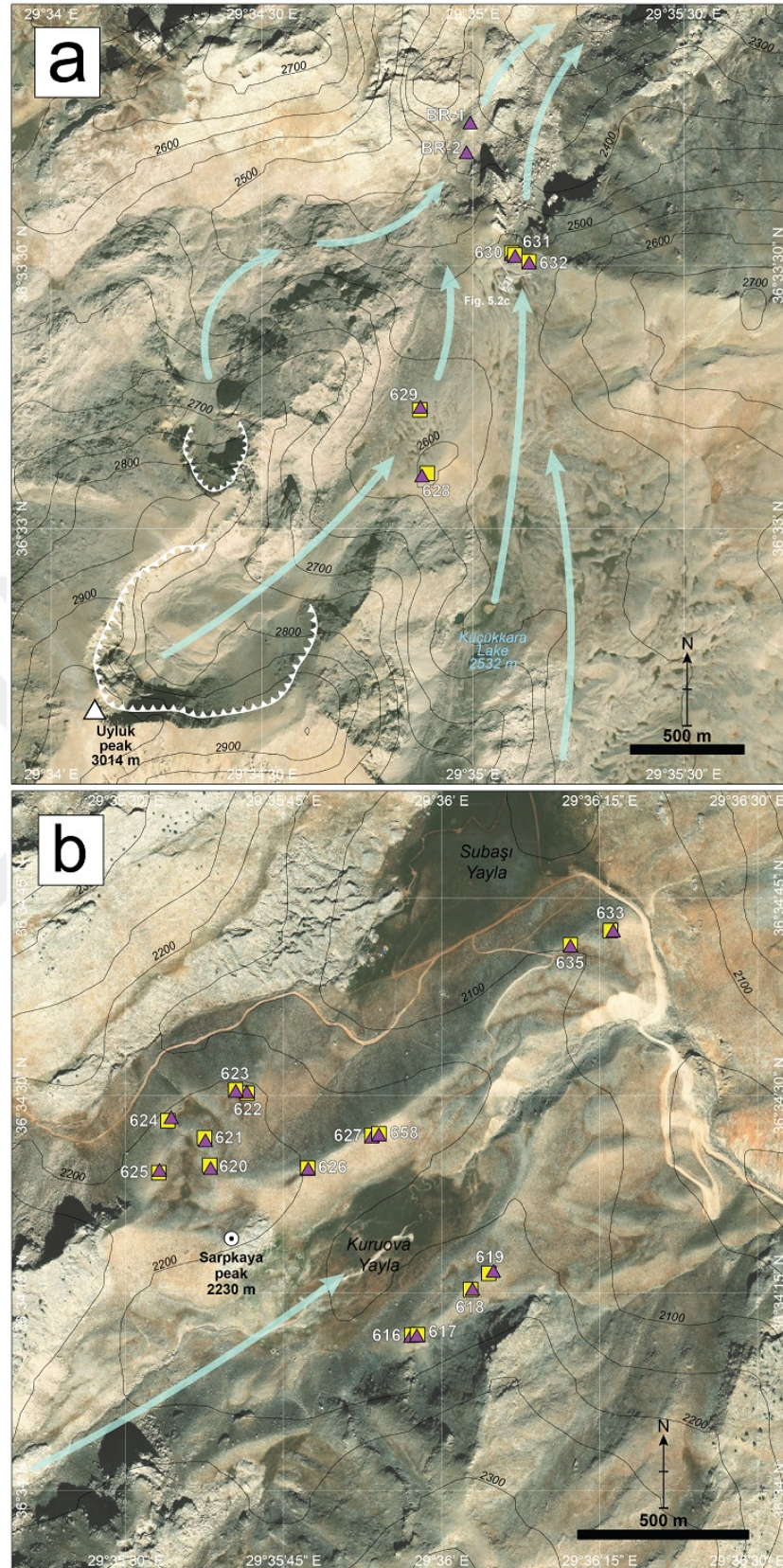


Figure 5.7 : The location of control and test boulders of Upper Kuruova (a) and Lower Kuruova valleys (b) in Akdağ. For clarity, only the numerical part of the Sample IDs is provided. The full Sample IDs are formatted as AK22-###-SH for control boulders and AK22-###-X for test boulders. Refer to Figure 5.4 for the legend.

There were no significant differences in the range of altitudes between the measured surfaces, except for Akdağ, which has a range of 500 m between 2085 and 2585 m a.s.l., due to the abundance of TCND samples from four relatively large-scale glacial valleys. On the other hand, the altitude range is 54 m (1893-1947 m a.s.l.) on Mt. Sandıras and 94 m (1945-2039 m a.s.l.) on Karadağ. In the selection of test surfaces, preference was given to those as close as possible to the control surface and with accessible horizontal or near-horizontal and flat microtopography for measurement. Mean R-values were calculated based on the 95% confidence interval of the sample size for each surface.

5.3.2 Generating calibration curves

In cases where a linear calibration between R-values and TCND ages is appropriate, a minimum of two CPs is necessary (Matthews & Winkler, 2022). In the initial phase of this study, a total of 58 CPs (Mt. Sandıras, $n=7$; Akdağ, $n=33$) were utilized to establish calibration curves. TCND outliers from both Mt. Sandıras ($n=2$) and Akdağ ($n=8$) are excluded from the calculations. Additionally, since the TCND ages of Karadağ have not yet been published (Bayrakdar et al., 2017; C. Bayrakdar, personal communication, June 5, 2024), a calibration curve could not be constructed using the R-value measurements taken here.

The calibration curves were generated using a linear regression model. This model forecasts the dependent variable (y) by considering the independent variable (x) through a linear relationship. The fundamental expression of the linear regression equation is as follows (5.1):

$$y = a + bx \quad (5.1)$$

In this equation, y represents the age of the CP (i.e., control surface) in years (dependent variable), x denotes the mean R-value of the CP (independent variable), while a and b are constants representing the intercept and slope of the calibration curve, respectively. The linear regression model is applied to the data using the least squares method, aiming to minimize the sum of squared variances between the observed and predicted values (5.2):

$$\sum_{i=1}^n (y_i - (a + bx_i))^2 \quad (5.2)$$

where n is the number of data points, (x_i, y_i) are the observed data points, b is the slope of the regression line, a is the y -intercept of the regression line.

5.3.3 Predicting SHD ages

After fitting the model, the coefficients a (intercept) (5.3) and b (slope) (5.4) are calculated:

$$a = \frac{\sum_{i=1}^n y_i - b(\sum_{i=1}^n x_i)}{n} \quad (5.3)$$

$$b = \frac{n(\sum_{i=1}^n x_i y_i) - (\sum_{i=1}^n x_i)(\sum_{i=1}^n y_i)}{n(\sum_{i=1}^n x_i^2) - (\sum_{i=1}^n x_i)^2} \quad (5.4)$$

These coefficients are essential as they define the angle and position of the line relative to the data points. Then, the model is capable of making predictions on new data points. By inputting a range of x -values, the model generates corresponding y -values based on the learned linear relationship, facilitating predictions beyond the observed data. Given a set of x values $x_1, x_2, \dots, x_n$ the predicted y values are obtained as (5.5):

$$\hat{y}_i = a + bx_i \quad (5.5)$$

In this way, based on the generated calibration curves, the SHD ages of the test surfaces were calculated. The consistency of the calculated ages is directly proportional to the negative relationship between the TCND age and the R -value of the CPs that form the calibration curve.

5.4 Results

The R -values obtained from glacial surfaces with similar TCND ages within the scope of the study exhibit differences based on lithological variations. For instance, average R -values measured from serpentinized harzburgite moraine boulders in Mt. Sandıras are much lower compared to the limestone lithology of Karadağ and Akdağ. The R -values in Mt. Sandıras appear to be much more precise compared to those of the other two mountains. Correspondingly, the highest R^2 values regarding the calibration

curves were obtained in Mt. Sandıras. On the other hand, a calibration curve that could be used to estimate SHD ages could not be generated in Akdağ. Consequently, it was not possible to create a calibration curve for Karadağ either.

5.4.1 R-values of control (SH) and test (X) boulders

The distribution of estimated mean R-values shows significant differences, even among similar lithologies. It was not found that there is a correlation between R-values and elevation, except in Mt. Sandıras ($r = 0.60$) (Figure 5.8a). However, the lowest R-values are also observed in Mt. Sandıras. Akdağ has the highest R-values. In the comparison of all R-values (both control and test surfaces) and R-values obtained only from control surfaces (i.e., CPs), except for Mt. Sandıras, no significant difference is apparent (Figure 5.8a and b). However, in the dataset where only the R-values of CPs are analyzed, a slight decrease is observed in the interquartile range (IQR) of each mountain (Figure 5.8b). The location with the lowest diversity of R-values appears to be Karadağ. The two outlier values observed in Akdağ belong to samples measured from glacially abraded bedrock surfaces (AK22-BR1-X and AK22-BR2-X).

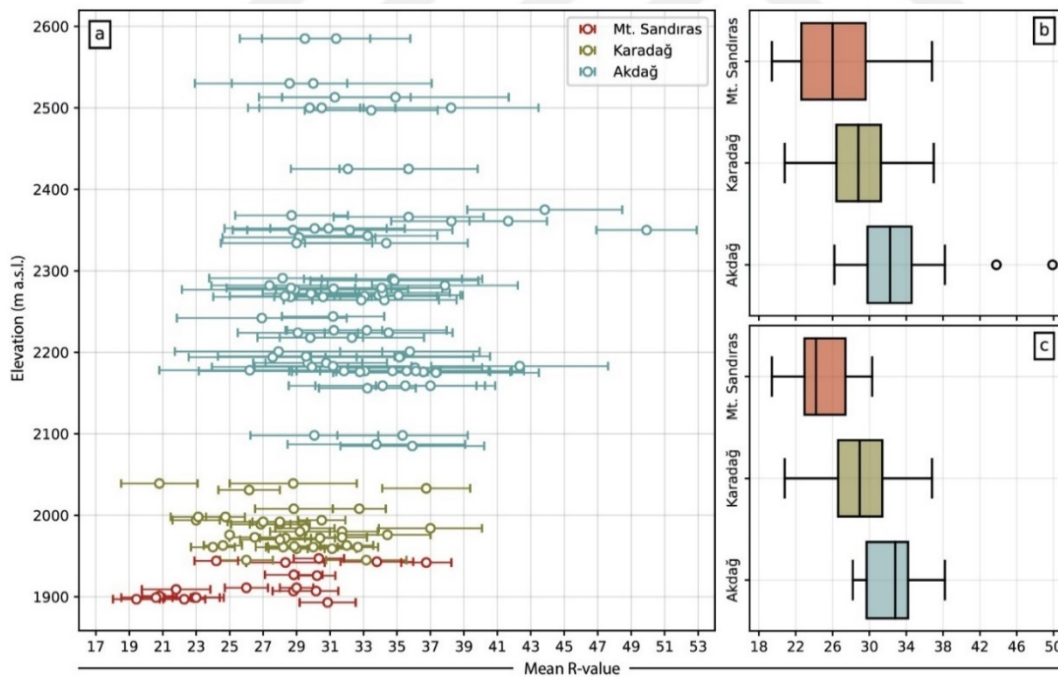


Figure 5.8 : R-values of sampled (control and test) surfaces. Elevation and R-values of the surfaces (a); box plot of control and test surfaces (b), and control surfaces (CPs) only (c). In the box plots, control points (CPs) within and younger than the Last Glacial Maximum (≤ 26.5 ka) have been used for comparison under the same conditions.

5.4.2 Calibration curves

At the outset, the calibration curves were generated by grouping the CPs based on (i) using all CPs (i.e., mountain variant calibration curves), (ii) the valleys where CPs are located (i.e., section-based calibration curves), and (iii) the glaciation periods indicated by the TCND age of CPs (i.e., time-variant calibration curves). The primary aim of this approach was to achieve the strongest negative relationship between TCND ages and R-values. However, most of these attempts failed and resulted in low coefficient of determination (R^2) values between the TCND ages and mean R-values of CPs. Only two calibration curves reject the null hypothesis (p -value < 0.05) (Figure 5.9a and 5.10a).

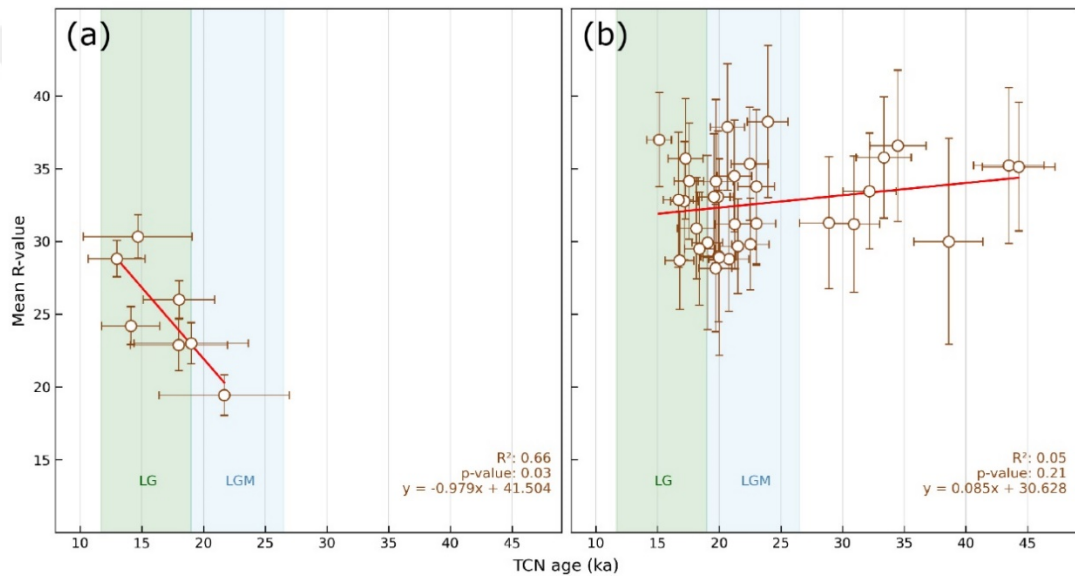


Figure 5.9 : Mountain-based calibration curves for Mt. Sandıras (a) and Akdağ (b). It is observed that the curve generated for Mt. Sandıras is much more consistent compared to that for Akdağ.

5.4.2.1 Mountain-based calibration curves

The variability within and between boulders and sites indicates lithological inconsistencies, microclimatic differences, and microtopographic irregularities, which are challenging to manage during field measurements (Matthews & Winkler, 2022). Accordingly, for the generation of calibration curves, the CPs were initially grouped based on the mountains (separately for Mt. Sandıras and Akdağ) (referred to hereafter as "mountain-based calibration curves"). This approach aimed to achieve consistent results in the calibration curves by considering both lithological and topographical differences. By limiting the sites in this way, the number of CPs used was kept at a

maximum (Mt. Sandıras, $n=7$; Akdağ, $n=33$). Consequently, a statistically significant calibration curve was obtained for Mt. Sandıras, although it was not sufficiently strong (R^2 : 0.66). On the other hand, the mountain-based calibration curve of Akdağ indicates no discernible fit ($R^2 = 0.05$) between the TCND ages and R-values (Figure 5.9b).

5.4.2.2 Section-based calibration curves

Due to the lack of expected results from the mountain-based calibration curves, valley-based CPs (referred to as section-based calibration curves) were utilized to identify the highest negative R^2 values between TCND ages and mean R-values of control surfaces. The highest result ($R^2 = 0.92$) was found with the CPs ($n=5$) located in the Kartal Lake valley (KLV) of Mt. Sandıras (Figure 5.10a). However, a calibration curve could not be generated for the Northwest valley of Mt. Sandıras due to the insufficient number of control surfaces ($n=2$). In Akdağ, no significant results were found based on the section-based calibration curves (Figure 5.10b-e). Among these calibration curves, the Karadere Valley (KV) CPs ($n=7$) exhibited the highest results ($R^2 = 0.30$), albeit low (Figure 5.9c). Moreover, it was statistically insignificant (p-value: 0.21).

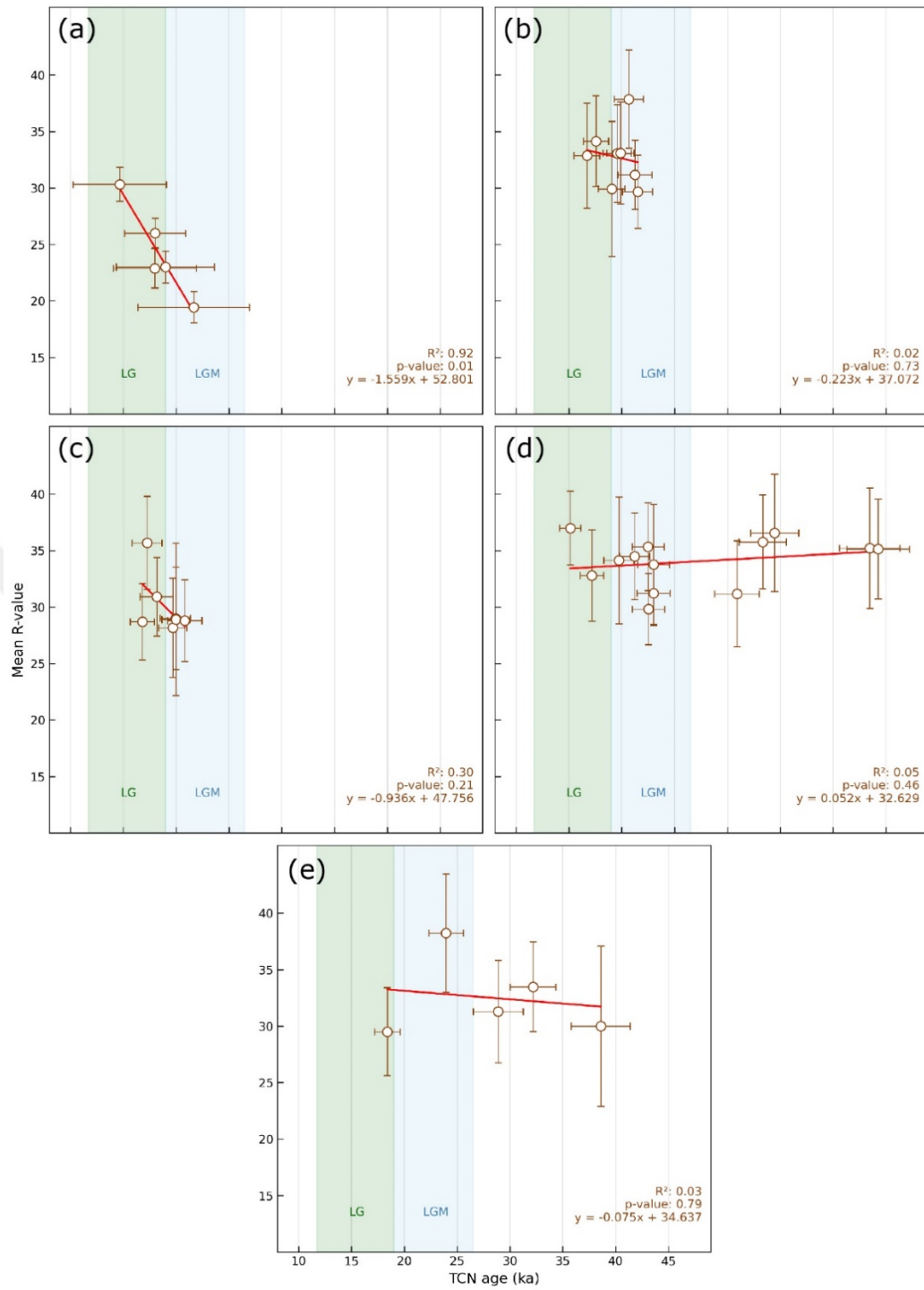


Figure 5.10 : Section-based calibration curves for Mt. Sandıras and Akdağ: (a) Kartal Lake Valley of Mt. Sandıras, (b) Taşkuzluklu Valley of Akdağ, (c) Karadere Valley of Akdağ, (d) Lower Kuruova Valley of Akdağ, and (e) Upper Kuruova Valley of Akdağ calibration curves.

5.4.2.3 Time-variant calibration curves

Based on the inconclusive results from the attempts to generate mountain and section-based calibration curves of Akdağ, time-dependent CPs were utilized to generate

"time-variant calibration curves". In this analysis, three distinct calibration curves were established based on the CPs within the time constraints of LGM-LG, LGM, and LG, respectively. All of these calibration curves were derived from Akdağ CPs due to the time variety of control surfaces. However, the calibration curves of Mt. Sandıras (Figures 5.9a and 5.10a) are also considered an LGM-LG calibration curve because of the TCND ages of its CPs. The time-variant calibration curves generated for Akdağ also yielded inadequate results. Beginning with the LGM-LG calibration curve, the aim was to exclude relatively older surfaces belonging to the pre-LGM period. However, the achieved results were unsatisfactory, showing no relationship between TCND ages and mean R-values of LGM-LG CPs ($n=25$) (R^2 : 0.00, Figure 5.11a). A similar situation was encountered in the LGM calibration curve of Akdağ, with a very low relationship between LGM CPs ($n=16$) (R^2 = 0.13, Figure 5.11b). Finally, the desired target was not achieved in the last calibration curve generated by LG CPs ($n=9$). However, it represents the highest result (R^2 = 0.42, Figure 5.11c) among the time-variant calibration curves of Akdağ.

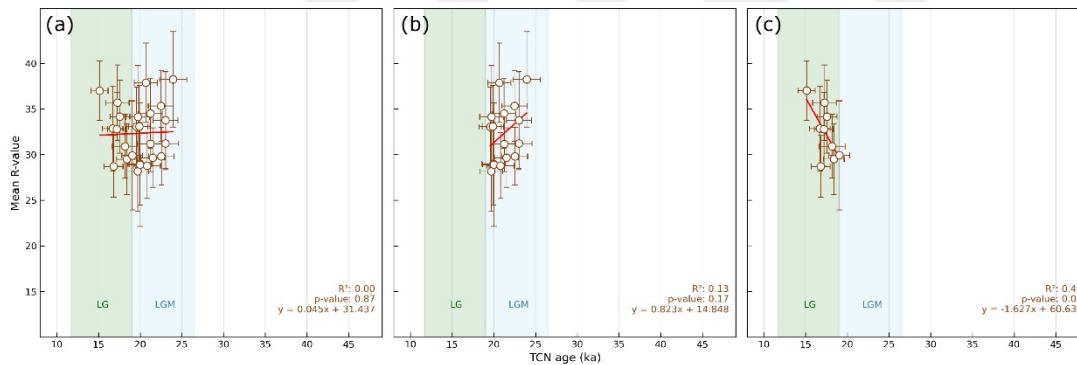


Figure 5.11 : Time-variant calibration curves for Akdağ: (a) LGM-LG, (b) LGM, and (c) LG calibration curves (LGM: Last Glacial Maximum; LG: Late-glacial).

5.4.3 Calculated SHD ages

Considering different combinations, only mountain and section-based calibration curves of Mt. Sandıras (Figure 5.9a and 5.10a) yielded statistically significant results (p -value: <0.05). Consequently, only the SHD ages of Mt. Sandıras were calculated (Figure 5.12 and Table 5.2).

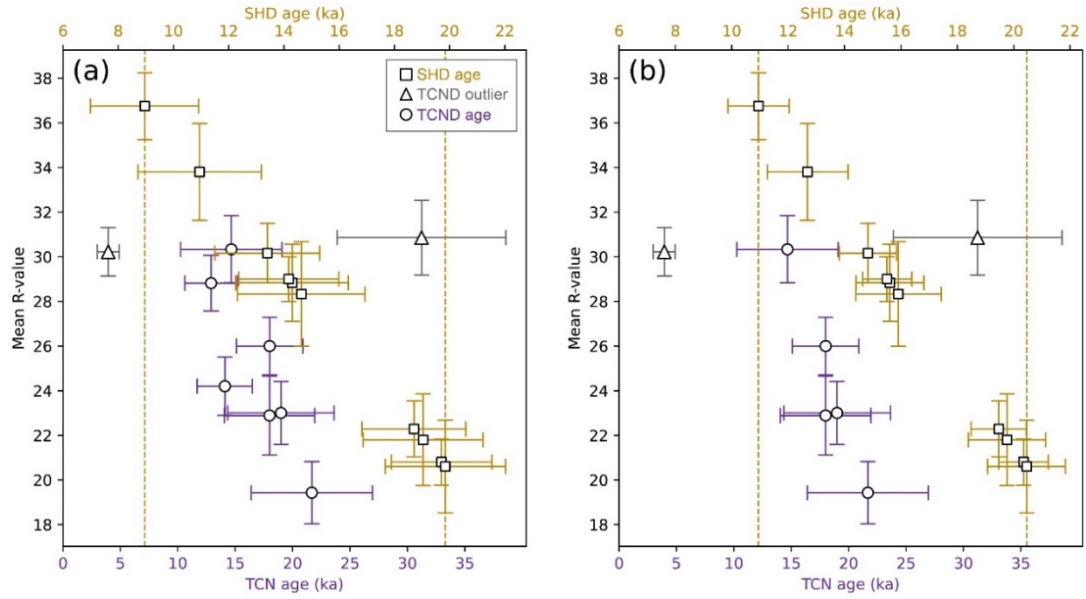


Figure 5.12 : TCND and SHD ages were determined for Mt. Sandiras. SHD ages were calculated using (a) all control points (i.e., the mountain-based calibration curve) and (b) control points from Kartal Lake Valley (section-based calibration curve), respectively. Dashed lines indicate the minimum and maximum calculated SHD ages on the secondary x-axis.

Based on the generated calibration curves, the SHD ages were calculated exclusively for Mt. Sandiras using both mountain and section-based calibration curves (Figure 5.12a and b). The calculation of the SHD ages shows differences between the results. For instance, the range of minimum and maximum values from the mountain-based calibration curve is slightly higher, and the standard deviations of the SHD ages are also higher based on these results (Table 5.2). Also, section-based calibration curve results have much higher minimum and maximum values (Figure 5.12b). It can be stated that the average difference between the two calculated age groups is approximately 1000 years.

Table 5.2 : Mean R-values with TCND and SHD ages of the moraine boulders in Mt. Sandıras. The TCND ages of the control boulders (-SH), along with the two different calculated SHD ages of the test boulders (-X), are shown separately (MBCC: mountain-based calibration curve; SBCC: section-based calibration curve). TCND outliers are indicated with an asterisk (*) at the end of the sample IDs.

Valley	Landform	Sample ID	Mean R-value	TCND age (ka)	SHD age (MVCC) (ka)	SHD age (SBCC) (ka)
Kartal Lake	A1 moraine	SA22-609-SH	19.4 ± 1.4	21.7 ± 5.3		
Kartal Lake	A1 moraine	SA22-609-X	22.3 ± 1.3		18.7 ± 1.9	19.5 ± 1.0
Kartal Lake	A1 moraine	SA22-610-SH	22.9 ± 1.8	18.0 ± 3.9		
Kartal Lake	A1 moraine	SA22-610-X	20.8 ± 1.0		19.7 ± 1.8	20.4 ± 0.9
Kartal Lake	A1 moraine	SA22-611-SH	23.0 ± 1.4	19.0 ± 4.6		
Kartal Lake	A1 moraine	SA22-611-X	20.6 ± 2.1		19.9 ± 2.2	20.5 ± 1.4
Kartal Lake	A2 moraine	SA22-612-SH	30.3 ± 1.5	14.7 ± 4.4		
Kartal Lake	A2 moraine	SA22-612-X	28.3 ± 2.3		14.6 ± 2.3	15.9 ± 1.5
Kartal Lake	A2 moraine	SA22-618-SH	26.0 ± 1.3	18.0 ± 2.9		
Kartal Lake	A2 moraine	SA22-618-X	29.0 ± 1.0		14.2 ± 1.8	15.5 ± 0.9
Kartal Lake	A2 moraine	SA22-619-SH*	30.9 ± 1.7	31.2 ± 7.3		
Kartal Lake	A2 moraine	SA22-619-X	21.8 ± 2.0		19.0 ± 2.2	19.8 ± 1.4
Northwest	B1 moraine	SA22-616-SH	28.8 ± 1.3	12.9 ± 2.3		
Northwest	B1 moraine	SA22-616-X	30.2 ± 1.3		13.4 ± 1.9	14.8 ± 1.0
Northwest	B2 moraine	SA22-613-SH*	30.2 ± 1.1	3.9 ± 1.0		
Northwest	B2 moraine	SA22-613-X	28.8 ± 1.7		14.3 ± 2.0	15.6 ± 1.2
Northwest	B2 moraine	SA22-617-SH	24.2 ± 1.3	14.1 ± 2.4		
Northwest	B2 moraine	SA22-617-X1	36.8 ± 1.5		9.0 ± 2.0	11.0 ± 1.1
Northwest	B2 moraine	SA22-617-X2	33.8 ± 2.2		11.0 ± 2.2	12.7 ± 1.4

Most of the control-test (SH-X) boulder ages are consistent with each other and signify reasonable results (Table 5.2 and Figure 5.13). In the Kartal Lake Valley, the TCND ages of samples #609, #610, #611 (hereafter “SA22-” will be denoted by “#”) (A1 moraine), and #618 (A2 moraine) are roughly clustered around the end of the LGM (~19 ka). The calculated SHD ages from the test boulders corroborate these TCND ages from the control boulders, except sample #618. However, among these samples, #609 and 618 appear to be problematic when comparing their mean R-value and TCND age with their SHD age. Sample #609 has the lowest mean R-value (19.4 ± 1.4) and the oldest TCND age (21.7 ± 5.3 ka) in Mt. Sandıras. Sample #612 and 618 (A2 moraine) represent the Late-glacial in Kartal Lake Valley (SHD ages: 14.6 ± 2.3 and 14.2 ± 1.8 ka).



Figure 5.13 : Certain control and test moraine boulders from Mt. Sandiras have both TCND and SHD ages. Sample IDs ending with "-SH" represent control boulders (i.e., control points), while those ending with "-X" denote test boulders. TCND outliers are indicated with an asterisk (*) at the end of the sample IDs. Double-headed white arrows indicate the control-test boulder sample groups.

Each valley contains a TCND outlier: sample #619 (TCND age: 31.2 ± 7.3 ka) (A2 moraine) in the Northwest Valley, and sample #613 (TCND age: 3.9 ± 1.0 ka) (B2 moraine) in the Kartal Lake Valley. However, their SHD ages (19.0 ± 2.2 ka and 14.3

± 2.0 ka, respectively) appear to be quite compatible with the ages of other samples located nearby in their respective valleys (Table 5.2 and Figure 5.4).

5.5 Discussion

5.5.1 Sampling design and strategy

In a study by Niedzielski et al. (2009), the required minimum sample sizes for certain lithologies were estimated based on the two-sample Student's t-test. They reported that the minimum sample size for limestones (e.g., Karadağ and Akdağ samples) falls within the range of 35-39. Although no analysis has been conducted on serpentinized harzburgite (e.g., Mt. Sandıras samples) surfaces within the same study, an analysis on serpentinite, demonstrates that setting the sample size at over 39 may yield more accurate results. They also emphasized the necessity of obtaining large samples (i.e., large sample size) for coarse-grained rocks like serpentinite, gabbro, gneiss, and greenstone. However, both calibration curves for Mt. Sandıras represent considerable results over a certain period (i.e., from LGM to Late-glacial). In other words, the sample size seems adequate considering the mountain and section-based calibration curves in Mt. Sandıras. On the other hand, a larger sample size might be more suitable for the calibration curves of Akdağ. In this manner, it might be possible to mitigate issues related to chemical weathering associated with carbonate rocks, thereby leading to more accurate results.

Surface irregularities are a common problem when obtaining SH measurements. To ease these issues, it is common to use a carborundum stone to prepare the test surface by abrading and smoothing the rock and removing loose particles for SH readings. It is called “pre-treatment” (cleaning) of the rock surface. Although this practice is necessary in disciplines such as rock mechanics, engineering geology, and related fields, it can be inconvenient for geomorphological studies such as surface weathering rates and exposure-age dating (Matthews & Winkler, 2022). In this study, although there was no need for pre-treatment on serpentinized harzburgite moraine blocks, such a situation was encountered with limestone blocks. Factors such as microtopography (e.g., microkarren formation), and the development of lichen and moss hindered the R-value measurements. It is believed that these challenges are one of the reasons why strong calibration curves could not be generated for Akdağ. Indeed, the presence of microkarren formations, such as rasp-like teeth features, minor rills, small rain pits on

many surfaces, and intense physical weathering (Figure 5.3), made it challenging to identify suitable surfaces for measurement. Moreover, relatively extreme lichen and moss development on certain limestone boulders is another reason to find proper surfaces for SH measurements.

5.5.2 R-value measurements

In the entire study area, notable differences in mean R-values are observed across both identical and disparate lithologies. The apparent disparity in R-values, notably the relatively lower values observed in Mt. Sandıras, appears to stem from the serpentized origin of the harzburgite boulders present in the vicinity. Despite sharing the same lithological composition, the mean R-values of Karadağ and Akdağ do not converge at a similar juncture. Considering the elevation of the samples from Karadağ and Akdağ (Figure 5.8a), this discrepancy may be linked to the deglaciation process experienced by the mountainous terrain. Under typical circumstances, the progression of a paleoglacier would follow a pattern from lower to higher elevations. Consequently, samples situated at lower elevations would be subjected to greater exposure to atmospheric conditions compared to their counterparts at higher elevations. Consequently, lower-elevation samples exhibit lower R-values. It can be observed that the samples from Karadağ generally have lower elevations compared to those from Akdağ (Figure 5.8a) and also exhibit a lower mean R-value interval (Figure 5.8b and c). While this observation does not provide solid evidence for a correlation between R-values and elevation in glacial limestone surfaces, it remains a minor detail that warrants consideration.

Glacially abraded bedrock surfaces measured in Akdağ exhibit the highest mean R-values within this study (AK22-BR1-X: 49.9 ± 3.0 ; AK22-BR2-X: 43.8 ± 4.6) (Figures 5.3; 5.8b and c). AK22-BR1-X is notable for its smooth and polished surface. In contrast, AK22-BR2-X features an irregular surface, similar to nearly all moraine boulders in Akdağ and Karadağ. This irregularity reflects the microkarren surface characteristics (e.g., rain pits) resulting from chemical weathering. Given their proximity (~100 meters), it is likely that the paleoglacier exposed AK22-BR1-X and AK22-BR2-X during similar periods. However, their mean R-values do not support this assumption. The difference in mean R-values between these two bedrock surfaces is likely attributable to the pre-debris cover on AK22-BR1-X, which was deposited

after the retreat of the paleoglacier. AK22-BR2-X extends as a flat surface, in contrast to AK22-BR1-X. This characteristic is likely the primary reason it has been (and continues to be) exposed to chemical weathering.

Indeed, SH measurements on bedrock surfaces tend to yield less variable R-values compared to boulder surfaces (Owen et al., 2007; Matthews & Winkler, 2022). Because, boulders may consist of mixed lithologies or possess unsuitable surfaces for measurement (Shakesby et al., 2006). Nevertheless, in the study area, all control points are moraine boulders. Although Mt. Sandıras presents reasonable R-values, the limestone surfaces of moraine boulders in Karadağ and Akdağ exhibit scattered results (Figure 5.8a). At this point, inhomogeneity and unfavorable surface features might be contributing factors. However, it is important to consider that limestone weathering may occur exponentially over time. Surface hardness may stabilize over time, resulting in similar R-value readings from surfaces with different exposure ages. For instance, post-glacial limestone surfaces younger than the Late-glacial period may yield more consistent results.

Although the temporal limitations of dating studies using SHD are generally confined to the Holocene, some studies have extended these limitations beyond the Holocene. In this context, the lithology of the study area is also important. For example, Tomkins et al. (2016) obtained consistent results on granite glacial surfaces older than 20 ka. In the study conducted as part of this thesis, two different lithologies are encountered: serpentinized harzburgite and limestone, which have a problematic structure, especially due to chemical weathering. Unlike Akdağ, Mt. Sandıras also has a temporal advantage. The oldest samples here date back to 21.7 ± 5.3 ka, while Akdağ has eight ages dating back to the pre-LGM. This is one of the reasons why a consistent calibration curve could not be obtained using all CPs (i.e., mountain-based calibration curve) at Akdağ. Additionally, as mentioned above, several problems related to chemical weathering have prevented the development of a consistent calibration curve(s) for Akdağ (Figure 5.9b, 5.10a-e, and 5.11a-c).

5.5.3 Consistency of R-values to establish a chronology

Despite its lower accuracy, the mountain-based calibration curve may provide more consistent results when considering the comprehensiveness of the predicted SHD ages (Table 5.2). With the higher error margins, it is believed that a more consistent

comparison can be made between the TCND ages and the SHD ages (Figure 5.9a and 5.10a). Based on this preference, the ages derived from the mountain-based calibration curve are considered appropriate for discussion in this section.

Even with the relatively larger difference between the TCND and SHD ages of sample #609, the A1 moraine in Kartal Lake Valley represents a consistent age group (Figure 5.4 and Table 5.2). Typically, it would be expected that the SHD age of sample #609 would also be the oldest among all samples. Interestingly, a similar discrepancy is observed even in the SHD ages calculated from the section-based calibration curve (R^2 : 0.92; Table 5.2). However, the A1 moraine likely provides the best evidence demonstrating that SHD is applicable to serpentinitized harzburgites. The SHD ages up to this point fully validate the TCND ages, providing a reasonable LGM chronology. The calculated ages of the two TCND outliers (samples #619 and #613) from Kartal Lake and Northwest valleys, show consistency with the other calculated ages in their respective valleys. Even though the mean R-value difference between the control and test boulders of sample #619 is high, the calculated age seems consistent with the general age group in A1 moraine of Kartal Lake Valley. Furthermore, the mean R-value harmony between the control and test boulders of sample #613 can be considered evidence for a Late-glacial age, similar to the other Late-glacial ages in B2 moraine of Northwest Valley.

According to the box plot analysis (Figure 5.8b and c), the mean R-values of the control and test surfaces of Mt. Sandıras differ from those of the samples from Karadağ and Akdağ. The overall mean R-value for the control and test surfaces (Figure 5.8b) in Mt. Sandıras is 26.4. Of the eight R-values higher than this mean, six are from test surfaces. The most problematic samples appear to be sample #617, 618, and 619. For each of these samples, there is either a control or test boulder with an R-value below the mean of 26.4 indicated in the box plot (Figure 5.8b). For instance, the test boulders of sample #617 (X1 and X2) exhibit unusual mean R-values compared to its control boulder (SH) and the rest of the data. These two test boulders may be part of a paraglacial rockfall that occurred on the slope southeast of the sample location (Figure 5.4). Based on the moraine grouping in the study by Sarıkaya et al. (2008), sample #618 is located in the A2 moraine along with sample #612 (Figure 5.4). However, their TCND ages (based on ICE-D Alpine data) differ significantly (18.0 ± 2.9 ka and 14.7 ± 4.4 ka, respectively). Conversely, their calculated ages are more consistent with each

other, being 14.2 ± 1.8 ka and 14.6 ± 2.3 ka, respectively. Considering the morphostratigraphy of the moraines, it seems reasonable to take into account the SHD age of sample #618. Finally, sample #619 (a TCND outlier, 31.2 ± 7.3 ka) stands out among the other test boulders with another significant SHD age. The mean R-value difference between its control and test boulder is also substantial (Table 5.2). This sample is located in the A2 moraine (Sarıkaya et al., 2008), which is possibly a Late-glacial moraine, particularly based on SHD age results. However, the calculated age of sample #619 is 19.0 ± 2.2 ka, aligning well with the A1 moraine, which dates back to the LGM.

5.6 Conclusion

The Schmidt Hammer Dating (SHD) method is both cost-effective and time-efficient compared to other, particularly numeric, exposure dating methods for dating various geomorphological processes. In this study, SHD was tested on glacial surfaces in a part of the Western Taurus Mountains. Unlike traditional applications of the method, a different approach was employed. A total of 137 moraine boulders and two bedrock surfaces were measured using an N-type Schmidt hammer, with 2780 impacts performed. According to the "control-test boulder strategy," while reasonable small-scale results can be obtained, many attempts to generate calibration curves from control points have failed. This suggests that lithology is crucial for obtaining meaningful results from numeric age–R-value comparisons and for generating reliable calibration curves.

The results from the serpentinitized harzburgite lithology of Mt. Sandıras were highly satisfactory and statistically significant based on the two generated calibration curves. Consequently, reasonable SHD glacial chronologies were established for Mt. Sandıras, marking the first SHD chronologies in Türkiye. Of the two chronologies established using different calibration curves, the most comprehensive one was used to evaluate SHD ages. In contrast, no relationship was found between the TCND ages and R-values of glacial surfaces in Karadağ and Akdağ.

The accepted SHD glacial chronology supports the two-phased deglaciation theory based on the TCND chronology in Mt. Sandıras. In the Kartal Lake Valley, the outer (A1) moraine represents a late LGM retreat, while the inner moraine just behind A1 indicates a Late-glacial retreat. In contrast to the two problematic test boulder ages in

the Northwest Valley, the SHD ages from the B1 and B2 moraines corroborate a Late-glacial retreat synchronized with the A2 moraine in Kartal Lake Valley. Although a few ages in the established SHD chronology of Mt. Sandıras appear inconsistent, the overall results are satisfactory.

In contrast to the positive outcomes observed in Mt. Sandıras, similar success was not achieved in Karadağ and Akdağ. Rational results regarding surface hardness could not be obtained, largely due to challenges associated with the chemical weathering of limestone surfaces. This weathering process affects surface hardness differentially and complicates the acquisition of R-values from samples by altering the microtopography of the surface. While serpentinization on harzburgite surfaces of Mt. Sandıras also appears to influence surface hardness, the extent of the issue is not as severe as observed in limestones. Additionally, the considerable age of the sampled surfaces, particularly dating back to the Last Glacial Maximum (LGM) and older periods, presented a significant challenge for both lithologies.

In prospective research endeavors, careful attention should be given to the time period relevant to the study. For instance, Holocene surfaces may warrant particular consideration, especially for glacial surfaces in limestone lithologies (or carbonate rocks in general). Instead of solely focusing on boulders, bedrock surfaces with TCND ages should be included in future studies. Furthermore, it is advisable to maintain a larger sample size whenever possible. Additionally, the utilization of machine learning methods may prove beneficial in addressing the challenges encountered in this study and similar contexts. Models capable of producing robust regression fits based on available R-value data can be developed not only for harzburgite and limestone but also for all lithologies presenting challenges in general.



6. CONCLUSION

In Chapter 2, the aforementioned glacial areas were identified based on their TCND chronologies. The ages were recalculated using up-to-date cosmogenic production rates and scaling schemas. In the Western Taurus Mountains, elevations are mostly below 3000 m a.s.l., and the Holocene-ELA of these mountains does not seem to be below their maximum elevations. Therefore, the absence of Late-glacial/Holocene ages is likely due to the relatively lower elevations of these mountains. Despite the much higher altitudes of the Eastern Black Sea Mountains (3900-4000 m a.s.l.), the absence of Late-glacial/Holocene ages might be attributed to the same reason.

Except for Erciyes Volcano, there are indications of Younger Dryas advances in the other mountains, particularly in Aladağlar. Generally, the Late-glacial to Holocene glaciers of Turkish mountains span from 14.8 ± 2.9 ka to 1.0 ± 0.3 ka, with some dating to the early to late Holocene. Evidence for glacier advances during the Little Ice Age (LIA) is also rare in Türkiye. However, it is suggested that rock glaciers may overtop these surfaces in upper valley systems, and there are also opinions regarding LIA surfaces on Erciyes Volcano.

Furthermore, despite the lack of Late-glacial/Holocene glaciers, there are still active glaciers, albeit in relatively small-scale areas. While there are valley glaciers, most exist as cirque glaciers located near the mountain peaks. The largest glacier in Türkiye is the ice cap on Mt. Ağrı, covering an area of 5.66 km². Additionally, Geyikdağ hosts the debris-covered glaciers in the westernmost part of the Taurus Mountains, which were recently discovered.

In Chapter 3, I presented the glacial geomorphology and timing of the Late Pleistocene deglaciations of Mt. Barla, the northernmost part of the Western Taurus Mountains. Mt. Barla has a distinct glacial history compared to other mountains in the region. First, I discussed the evolution of glacial valleys based on polygenetic and polycyclic processes. While glaciokarst (a common phenomenon, especially around the Mediterranean) controls the development of the Gedik and Ortayazı valleys, the

Kapıdere Valley has a notably complex development history. Mass movements appear to have been the initial trigger for glacial processes in this valley.

Subsequently, there were at least four glacial retreats, evidenced by four recessional moraines that I mapped. The Kapıdere Valley, though modest in geomorphology and smaller in scale compared to other valleys on Mt. Barla, appears to be the most dynamic valley in the context of Late Pleistocene glaciations.

The Late Pleistocene glaciations in Mt. Barla date back to 78.5 ± 17.6 ka (MIS 5), based on the weighted average ages of two TCND samples from a lateral moraine in the Kapıdere Valley. However, this claim requires additional evidence, as the lower part of the valley has been significantly deformed by postglacial processes. In the same area, two terminal moraine ages indicate 45.5 ± 6.8 ka (MIS 3), which correlates with an individual terminal moraine age of 42.1 ± 5.2 ka from the Gedik Valley. Finally, the oldest landform age of 16.6 ± 2.6 ka designates Late-glacial deglaciation in the Gedik Valley.

Chapter 4 focuses on establishing glacial chronologies in the Western Taurus, specifically on Mt. Davraz. Although it is in close proximity to Mt. Barla, Mt. Davraz exhibits a completely different glacial development. Observations indicate that piedmont-type glaciations shaped the north-facing slopes of the mountain. I mapped 10 cirques aligned in an E-W direction, all with a roughly northern aspect. These cirque glaciers served as the source for the piedmont glaciers. The winds from the south significantly influenced the strong cirque-type glaciation, as they prevented glacier development on the high plateau-like top of Mt. Davraz.

Interestingly, there are no well-developed glacial valleys, which are typically common in piedmont-type glaciations. As a result, the cirque glaciers moved directly through the piedmonts and extended onto a relatively flat area. Once the glaciers disappeared, hummocky moraines were exposed on the surface. According to six TCND samples from these hummocky moraines, the glaciers began to retreat by 21.8 ± 2.4 ka (LGM) from the east of the hummocky moraine field. The glaciers completely disappeared by 17.7 ± 2.2 ka (Late-glacial).

In Chapter 5, the final part of this thesis, I employed the Schmidt hammer exposure-age dating (SHD) method on Mt. Sandıras, Karadağ, and Akdağ. While the results were not entirely satisfactory, they did provide notable insights based on the

application of the method. Unlike traditional approaches, I employed a slightly different methodology to test SHD on glacially abraded harzburgite (Mt. Sandıras) and limestone (Karadağ and Akdağ) surfaces. For each control point (boulders with TCND ages), I measured the R-values of test boulders (boulders without any ages).

The results from the harzburgites were reasonably positive, with the SHD ages showing considerable alignment with the TCND ages of Mt. Sandıras. Additionally, this method allowed for the evaluation of TCND outliers based on the R-values of the test boulders, providing a secondary opportunity to interpret the chronology. Specifically, I found that the SHD ages of two outliers from Kartal Lake and the Northwest valleys were consistent with their surrounding areas.

In contrast, the samples from the limestone lithologies of Karadağ and Akdağ yielded completely chaotic results. The primary issue was the negative impact of chemical weathering on measuring the surface hardness of the rock, which made it challenging to obtain accurate R-values due to the microkarren surface characteristics. Even when measurements were possible, the surface hardnesses varied significantly. The only notable conclusion for limestone was that many samples from Karadağ had a lower average R-value compared to those from Akdağ. Most of these samples were located between 1950-2000 meters above sea level, while the Akdağ samples were situated at elevations roughly above 2100 meters.

With this thesis, I presented new glacial chronologies and took the initial step toward implementing an untested methodology for the Western Taurus Mountains. Terrestrial Cosmogenic Nuclide Dating (TCND) is a powerful yet challenging method for revealing past geomorphological processes. However, it remains quite expensive. On the other hand, methods like Schmidt hammer exposure-age dating (SHD) are proven to be time- and cost-effective, though they are not compatible with most lithologies.

As a well-established methodology, TCND will undoubtedly continue to be refined and developed. However, SHD also has significant potential for growth. I believe that integrating technologies such as artificial intelligence and machine learning, which can enhance both the sample collection and analysis processes, will significantly advance the SHD methodology.



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APPENDICES

APPENDIX A: Additional information about Chapter 5 (Schmidt hammer exposure-age dating).





APPENDIX A

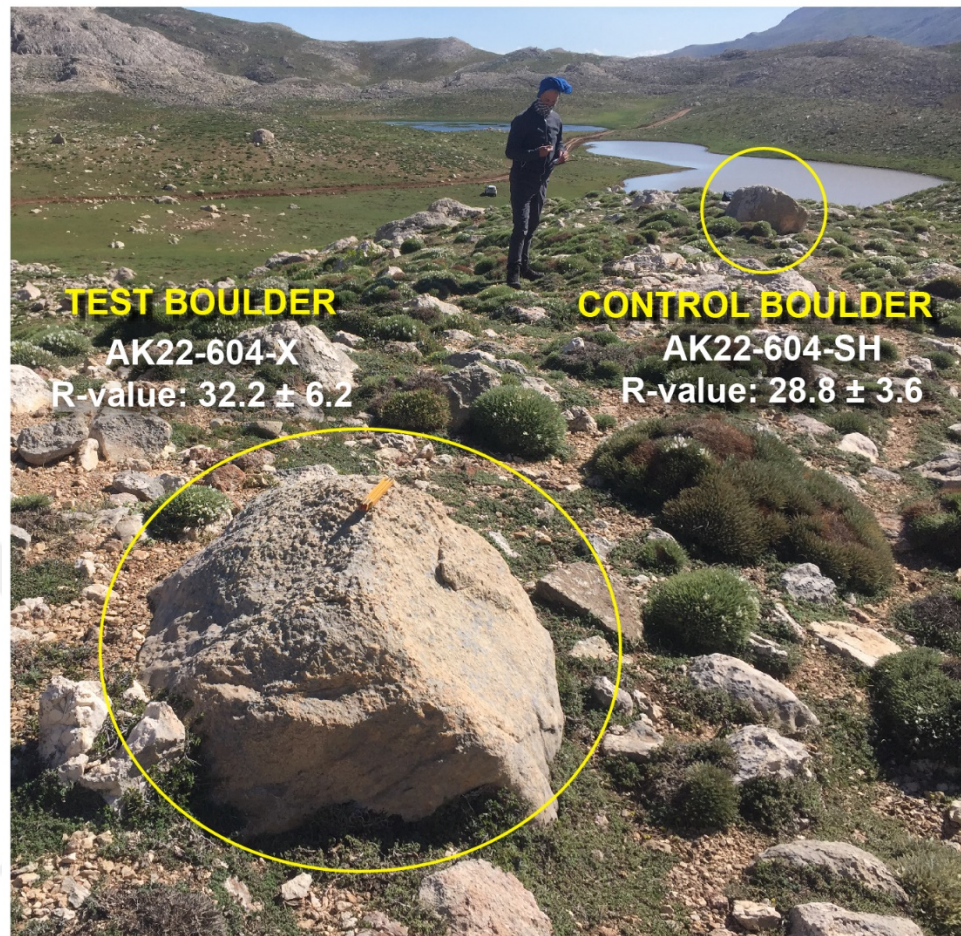
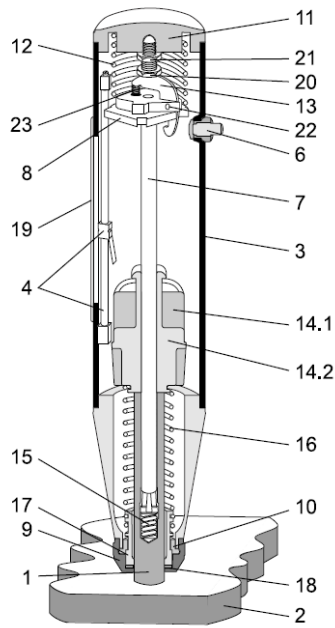


Figure A.1 : A control-test boulder group from the right lateral moraine of the Karadere Valley, Akdağ.

3.4.4 Concrete Test Hammer Model N/L



Key:

- 1 Impact Plunger
- 2 Test surface
- 3 Housing
- 4 Rider with guide rod
- 5 Not used
- 6 Push button, complete
- 7 Hammer guide bar
- 8 Guide disk
- 9 Cap (for models L and LR red)
- 10 Two-part ring
- 11 Rear Cover (for models L and LR red)
- 12 Compression spring
- 13 Pawl
- 14 Hammer mass: 14.1 model N, 14.2 model L
- 15 Retaining spring
- 16 Impact spring
- 17 Guide sleeve
- 18 Felt washer
- 19 Plexiglas window
- 20 Trip screw
- 21 Lock nut
- 22 Pin
- 23 Pawl spring

Fig. 3.3 Lengthways section through the concrete test hammer

4 Data

4.1 Form of Delivery

Concrete test hammer	Model N	Model NR	Model L	Model LR
Article-no.	310 01 000	310 02 000	310 03 000	310 04 000
Total weight	1,7 kg	2,6 kg	1,4 kg	2,4 kg
Carrying case, W x H x D	325 x 125 x 140 mm	325 x 295 x 105 mm	325 x 125 x 140 mm	325 x 295 x 105 mm
Grindstone	1 pce.	1 pce.	1 pce.	1 pce.
Recording paper	–	3 rolls	–	3 rolls

4.2 Accessoires

Concrete test hammer	Model N	Model NR	Model L	Model LR
	Accessory article number			
Testing anvil	310 09 000	310 09 000	310 09 000	310 09 000
Recording paper, pack of 5 rolls	–	310 99 072	–	310 99 072

4.3 Technical Data

Concrete test hammer	Model N	Model NR	Model L	Model LR
Impact energy	2,207 Nm		0,735 Nm	
Measuring range	10 to 70 N/mm ² resistance to pressure		10 to 70 N/mm ² resistance to pressure	

Figure A.2 : The general specifications of the Proceq N-type Schmidt hammer used in this study (from Proceq, 2004).



Figure A.3 : Measuring the marble calibration boulder located in front of the Faculty of Mines, Istanbul Technical University, Turkey (left), and a control boulder (SA22-617-SH) from the B2 moraine of the Northwest Valley, Mt. Sandıras (right).

Table A.1 : SH measurements of the İTÜ calibration boulder

June 10, 2022				June 29, 2022			
Impact #	R-value			Impact #	R-value		
1	54			1	49		
2	59			2	50		
3	59			3	50		
4	59			4	50		
5	60			5	50		
6	60			6	50		
7	60			7	50		
8	60			8	51		
9	60			9	51		
10	61			10	51		
11	61			11	51		
12	61			12	51		
13	61			13	52		
14	61			14	52		
15	61			15	52		
16	62			16	52		
17	62			17	52		
18	62	Mean R-value		18	53	Mean R-value	
19	63	Average	Standard deviation	19	54	Average	Standard deviation
20	63	61.7	0.8	20	54	52.3	1.1
All-->		60.5	1.9	All-->		51.3	1.4
95% conf. int.		60.5	0.5	95% conf. int.		50.5	0.5



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