Chapter 15
The Carbon Cycle in Drylands

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Abstract Drylands are characterized by an aridity index (ratio of annual rainfall to potential evapotranspiration) lower than 0.65, and occupy nearly a third of the total land surface. Globally, the organic and inorganic carbon (C) storage in such water-limited systems is about 20–30% of the terrestrial global total. The total soil organic C (SOC) stored in drylands is approximately 230 Pg. The C content in dryland biomass is about four times lower than that stored as SOC (65 Pg). The soil inorganic C (SIC) pools are estimated to be more than twice the SOC pools for drylands and may exceed SOC by a factor of 10 in some arid lands. These statistics can be modified significantly taking into account anthropogenic practices. Ideally, NT management may potentially increase the SOC by 20%, while non-grazing in grassland could increase SOC storage by about 45%. These ecosystems are highly vulnerable to climatic changes and susceptible to desertification, leading to reduction in the C pool. In addition, due to arid conditions and the large percentage of bare soil, some other processes besides photosynthesis and respiration contribute to C sequestration or gaseous emissions to the atmosphere. These include geochemical

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R. Lal et al. (eds.), *Recarbonization of the Biosphere: Ecosystems and the Global Carbon Cycle*, DOI 10.1007/978-94-007-4159-1_15,

processes, formation of secondary carbonates, bio-sequestration, subsoil ventilation, erosion and photodegradation and can even dominate the ecosystem C exchange during the dry season. Such contributions limit the use of biological models to provide estimates of C pool in drylands. And the differences in the measured net C exchange with the atmosphere, ranging from −106 to 145 gC m^{-2} for deserts and from −190 to 140 gC m^{-2} for grasslands are not easily explained. A better understanding of C cycle in drylands is highly relevant to the Kyoto Protocol in order to prevent degradation and the C emissions to the atmosphere. This survey reviews measured C pools and annual C sink capacities in drylands, together with an analysis of principal processes involved and, finally, summarizes suggestions for management practices with the capability to reduce C losses from drylands.

Keywords Drylands • Pools • Desertification • Geochemical processes • Ventilation • Soil inorganic C pool • Weathering • Erosion • Biosequestration • Photodegradation • Net erosion exchange • Global C cycle • Aridity index • Evapotranspiration • Ventilation • Anthropogenic emissions • Missing sink • Terrestrial ecosystems • Mitigation • Photosynthesis process • Rangeland • Desertification • Arid • Hyper-arid • Desert biomes • Caliche • No-till • Minimum tillage • Lithogenic carbonates • Primary carbonates • Secondary carbonates • Pedogenic carbonates • Silicate weathering • Calcite • FLUXNET • Flux measurements • Carboniferous rocks • Biomineralization • Phytoliths • Eddy covariance

Abbreviations

P	annual rainfall
P/E_P	aridity index
C	carbon
CO_2	carbon dioxide
GCC	global carbon cycle
NEE	net ecosystem CO_2 exchange
E_P	potential evapotranspiration
SIC	soil inorganic C
SOC	soil organic C
UV	Ultraviolet

15.1 The Global Carbon Cycle

The global carbon (C) cycle (GCC) depends on feedbacks among a number of source and sink processes occurring among different systems: ocean, atmosphere, soil and biosphere. These processes operate at different time scales modifying the C composition of components (Boucot and Gray 2001). In the last decades, the increase of atmospheric C via anthropogenic carbon dioxide (CO_2) emissions and changes in land use has produced a climatic perturbation inducing changes in temperature and rainfall regimes (IPCC 2007; Keeling 1960). Moreover, the effect of such perturbations may also be altering other systems beside the atmosphere such as the soil and biosphere. The annual increase of the atmospheric CO_2 concentration is only half that expected from anthropogenic activities, implying a terrestrial or oceanic sink absorbing CO_2. Isotopic studies reveal that air–sea CO_2 exchange is too small to explain the "missing sink" which must, therefore, be accounted for by terrestrial ecosystems (Schimel et al. 2001; Tans et al. 1990). Thus, a better understanding of the role of the biosphere in the current global C budget as well as the potential of soil as a C storage medium is needed in order to enable the mitigation of human impacts.

The GCC is strongly related to the C balance of terrestrial ecosystems due to the capacity of the biomass and soil to store C. The biosphere, via the photosynthetic process, captures CO_2 from the atmosphere and stores it in the living biomass. Then, soil microorganisms degrade the non recalcitrant compounds of dead biomass emitting CO_2 to the atmosphere. The organic matter can be stabilized and stored in the soil at long time scales via spatial inaccessibility to decomposer organisms or interactions with minerals and metal ions (von Lützow et al. 2008). Thus, the soil is the largest pool of organic C in terrestrial ecosystems, representing a reserve of more than 1,500 Pg C (1 Petagram = 10^{15} g) (Safriel et al. 2005). In addition, soil contains more than 900 Pg C in inorganic forms such as calcite or dolomite (Safriel et al. 2005; Vande Walle et al. 2001), while living biomass represents a C reserve of about 600 Pg. Therefore, terrestrial ecosystems with deeper soils and greater biomass, such as forests, present higher potential to sequester CO_2 from the atmosphere and contribute to amelioration of anthropogenic CO_2 emissions and, thus, climate change.

15.2 Main Characteristics of Drylands

Drylands are characterised by patches of vegetation and bare soil exposed to erratic rainfall events producing water-stressed vegetation during the drought period (Domingo et al. 1999). These water-limited ecosystems exist on every continent and comprise nearly a third of the total land surface corresponding to 60 million km^2 (Okin 2001; Schlesinger 1990). Dryland rangelands support about 50% of the world's livepool and provide forages for both domestic animals and wildlife (Puigdefábregas 1998). Although drylands withstand extreme climatic conditions, they are very sensitive to perturbations such as drought, fires or climate change, leading to desertification (Mouat and Lancaster 2006). This process can be defined as land degradation in arid, semi-arid, and dry sub-humid areas resulting from climatic variations and human activities (UNEP 1997). Desertification reduces the potential for plant C assimilation, degrades soil and, thus, decreases dryland C pools. One of the major trends in the degradation of dryland ecosystems is the replacement of grass by shrubs (Puigdefábregas 1998). This trend alters soil properties and can modify the amount of C stored in biomass and soil.

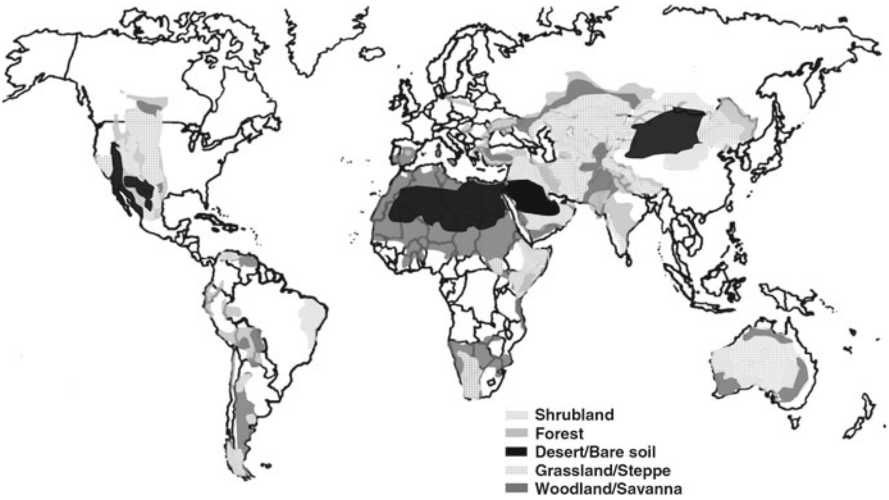

Fig. 15.1 Global distribution of main ecosystem types in drylands

According to the World Atlas of desertification (UNEP 1997), a terrestrial ecosystem is considered a dryland when the ratio of annual rainfall (P) to potential evapotranspiration (E_p), i.e., the aridity index (P/E_p), is lower than 0.65. Drylands are divided into four subtypes, in order of increasing aridity or moisture deficit: dry sub-humid (11% of total dryland surface), semi-arid (31%), arid (34%) and hyper-arid (24%). Desert biomes or bare soil predominate in hyper-arid and arid subtypes, while grasslands typify semi-arid, and woodland or forest survive in dry sub-humid drylands; about 35% and 40% of semi-arid and dry sub-humid subtypes, respectively, are cultivated (mainly croplands) (Safriel et al. 2005, Fig. 15.1).

Generally, drylands represent a limited potential to store C due to their poor soils and sporadic and dispersed vegetation. However, their large extent around the terrestrial land surface explains the estimated soil organic C (SOC) pool content of about 400 Pg C, and a contribution of about 20–30% to the terrestrial organic and inorganic C (Eswaran et al. 2000; Rasmussen 2006; Safriel et al. 2005). In this context, the main characteristic of drylands is the capacity to store about 95% of the soil inorganic C (SIC) globally via “caliche formation” (Marion et al. 2008). However, the role of the SIC pool in relation to climate change is less well understood (Lal and Kimble 2000b). Most carbonate rocks are found in drylands (mainly in the Mediterranean, China and Africa) where other processes besides photosynthesis and respiration (biological processes) also contribute to C sequestration or C emission to the atmosphere (Serrano-Ortiz et al. 2010).

This chapter of the C cycle in drylands reviews measurements of C pools and annual C sink capacities together with analysis of the main drivers controlling the potential C sequestration capacity and principal processes involved. Finally, some suggestions for management practices to improve the role of drylands in the storage of SOC are summarized.

15.3 Carbon Pools

15.3.1 Soil Organic Carbon

Several studies have been published related to SOC pools in drylands (Table 15.1).

According to published information regarding SOC pool in some studied deserts and bare soil, it ranges from 1 to 3 kg C m^{-2} in Africa and North and Central of America (Rasmussen 2006; Woomer et al. 2004), while it ranges from 2 to 7 kg C m^{-2} in Asia (Wiesmeier et al. 2011). For shrublands, SOC ranges from 2 to 4 kg C m^{-2} except on the Asian continent where SOC higher than 6 kg C m^{-2} has been reported (Chen et al. 2007; Wiesmeier et al. 2011). The average value of SOC in grassland is 5 ± 2 kg C m^{-2}, but can be more than double for non-grazed systems (He et al. 2008). However, there is little information for woodland or savannas. Some studies have reported a range of about 3–5 kg C m^{-2} (Chen et al. 2007; Noellemeyer et al. 2006; Shukla et al. 2006). Forests are not a common vegetation in drylands. For SE Spain, Martínez-Mena et al. (2008) reported a value of 14 kg C m^{-2} while a study in Kenyan dryland forest indicated an average value of 2.3 kg C m^{-2} mainly due to differences in climate conditions and species (Glenday 2008). About 30% of drylands are cultivated. Dry croplands have an average SOC pool of 3.5 kg m^{-2}. These are mostly located in dry-subhumid and semiarid ecosystems, and store globally about 40 Pg SOC. Many studies suggest an improvement in C sequestration if non-tillage (NT) or minimum-tillage practices are used instead of conventional ploughing (López-Fando and Pardo 2009, 2011; Sombrero and de Benito 2010).

Globally, about 230 Pg of total SOC is stored in drylands according to the coverage of drylands per continent occupied by the different ecosystem types (Fig. 15.1), with average SOC values summarized in Table 15.1. The average SOC is in the same range of values published by Lal (2004) (241 Pg C) but higher than those published by IPCC (1990) and Bolin et al. (2001) (191 and 159 Pg C, respectively). However, these estimates are highly uncertain mainly because of low sample numbers used for global upscaling and assumptions on mean soil depths (Rodeghiero et al. 2009). In addition, global values can be easily modified considering anthropogenic practices such as taking into account the percentage of cultivated lands (SOC reduced by 15%), an ideal situation of NT which would increase the total SOC in drylands by 20% (258 Pg C), and by prohibiting grazing in grassland which could increase the value by about 45% (313 Pg C).

15.3.2 Soil Inorganic Carbon

The SIC pool consists of primary inorganic carbonates or lithogenic inorganic carbonates, and secondary inorganic carbonates or pedogenic inorganic carbonates (Saharawat 2003). Extra inputs of Ca^{2+} due to atmospheric deposition and/or silicate weathering combined with the negative water balance may result in calcite precipitation

Table 15.1 Soil organic carbon (kg C m^{-2}) from different dryland ecosystems together with mean annual temperature and precipitation

Continent	Location	Reference	Mean annual temperature (°C)	Mean annual precipitation (mm)	Soil depth (cm)	Vegetation	Management	SOC (kg C m^{-2})
Europe	SE Spain	Oyonarte (personal communication)	18	200	0–50	Cropland	–	3
						Shrubland	–	3
						Grassland	–	4.5
		Martínez-Mena et al. (2008)	16.6	300	–	Forest	–	14
						Abandoned	–	8.2
						Olive	Non irrigated	7.2
	NE Spain	Plaza-Bonilla et al. (2010)	–	430	0–40	Cropland	No Tillage	3–3.5
	Center Spain	López-Fando and Pardo (2009, 2011)	–	400	0–30	Cropland	No Tillage	5
							Conventional Tillage	4
		Sombrero and de Benito (2010)	–	448	0–30	Cropland	Conventional Tillage	4
							Minimum Tillage	5
							No tillage	10
Asia	Center China	Chen et al. (2007)	34 – (–27)	427	0–40	Cropland	–	3
						Grassland	–	5
						Shrubland	–	6
						Woodland	–	4
	North China	He et al. (2008)	1.1	345	0–100	Grassland	Grazing	6–14
		Wang et al. (2009)	(–19) – (–23)	350	0–30	Cropland	28 year cropland	1
							42 year cropland	2
		Wiesmeier et al. (2011)	0.7	350	0–100	Arable	–	11 ± 4
						Bare	–	5 ± 2
						Steppe	–	14 ± 5
						Sand Dunes	–	5 ± 3
	India	Singh et al. (2007)		100–400	0–100	Bare	–	2.4

America	Mexico	Shukla et al. (2006)	15	400	0–20	Oak	–	8
						Juniper	–	4
	USA (Arizona)	Rasmussen (2006)	–	–	–	Arid	–	1.4–2.8
		Emmerich (2003)	17	356	0–30	Grassland	–	2.6 ± 0.5 (Spring) 2.3 ± 0.1 (Fall)
						Shrubland	–	4.0 ± 0.5 (Spring) 2.9 ± (0.3) (Fall)
	USA (Nebraska, Colorado)	Denef et al. (2008)	9.5	500	0–75	Native grassland	–	5–9
						Dryland cultivation	–	6–7
						Pivot Irrigated	–	7–8
	Argentina	Noellemeyer et al. (2006)	16	480	0–18	Grassland	–	1
						Grass + shub + trees	–	2.7
	Chile	Perez-Quezada et al. (2011)	26–5	153	0–50	Shrubland	–	2–4
		Muñoz et al. (2007)	–	695	–	Afforested Shrubland	–	4.2
Africa	Sahel desert	Woomer et al. (2004)	–	–	0–40	Desert	–	2
	Kenya	Glenday (2008)	–	–	–	Forest	–	2.3
	Tanzania	Birch-Thomsen et al. (2007)	20	542	0–50	Maize cultivation	–	2–4

forming secondary carbonates (caliche) and contribute to SIC sequestration (Marion 1989; Schlesinger 1985). The contribution of inorganic C formations from non-carbonate material (caliche) may range from 0.12 to 0.42 g C m^{-2} $year^{-1}$ (Marion et al. 2008; Schlesinger 1985). Since SIC is relatively stable, with turnover periods >1,000 years (Amundson et al. 1994), the C stocks in soil are generally similar following land use and management changes and it is usually not considered in soil C dynamics (Allen et al. 2010). What is more, inorganic C formed from re-precipitation of calcareous material may not be involved in C sequestration in the soil.

The SIC pools are estimated to be more than twice the SOC pools for drylands (Eswaran et al. 2000; Lal and Kimble 2000a). Further, SIC pools may exceed SOC by a factor of 10 in some arid lands (Schlesinger 1985, 2006). What is more, a study in a site located in Southeasth Spain reveal an average of SIC pool 17 times that of SOC content (134 kg m^{-2}) (Díaz-Hernández et al. 2003). Since SIC Although, there is no clear evidence to confirm an effect of SIC on SOC, soils with caliche formations are almost twice as rich in SOC as in those of a similar depth but without such horizons (Díaz-Hernández and Barahona Fernádez 2008). Soils with high SIC are mainly located in hyper-arid and arid regions with a pool of about 732 Pg C (Safriel et al. 2005). Pools of SIC in semi-arid and dry sub-humid systems are almost four times lower.

15.4 Biomass Organic Carbon

The C content in dryland biomass is about four times lower than that stored as SOC (Eswaran et al. 2000). The vegetation is mainly comprised of grass, steppe and woody species with a large proportion of bare soil and, thus, a low capacity to store C. Hyper-arid lands (deserts) in Asia and Africa have the capacity to store 0.04–0.40 kg C m^{-2} in biomass (Fan et al. 2008; Woomer et al. 2004). Biomass C storage in shrublands is in the range of 0.08–0.40 kg m^{-2} depending on the percentage of bare soil and the degree of degradation (Perez-Quezada et al. 2011). The C content in woody species, mostly located in Africa, ranges from 0.9 to 2.6 kg C m^{-2} depending on species and climatic conditions (Shackleton and Scholes 2011; Williams et al. 2008). Although forest occupies less than 15% of drylands, its capacity to store C can be about 4–5 kg C m^{-2} (Glenday 2008). Finally, grasslands located mainly in Asia store around 1 kg C m^{-2} with decreased potential depending on the grazing intensity (He et al. 2008).

Globally, average pool of 65 Pg of total C in the biomass is estimated based on the percent of land cover (Figs. 15.2 and 15.3) and average value of published C content of biomass during the last 5 years. This value can be modified significantly considering some anthropogenic practices such as grazing, which may reduce C pools by more than 10% of the given value. Degradation of the vegetation comprised of woody and shrubland types due to desertification may reduce C pool in the biomass by more than 20% (50 Pg C). The estimated potential biomass C based on the maximum estimated values is about 81 Pg C, and this is in accord with that reported by Safriel et al. (2005).

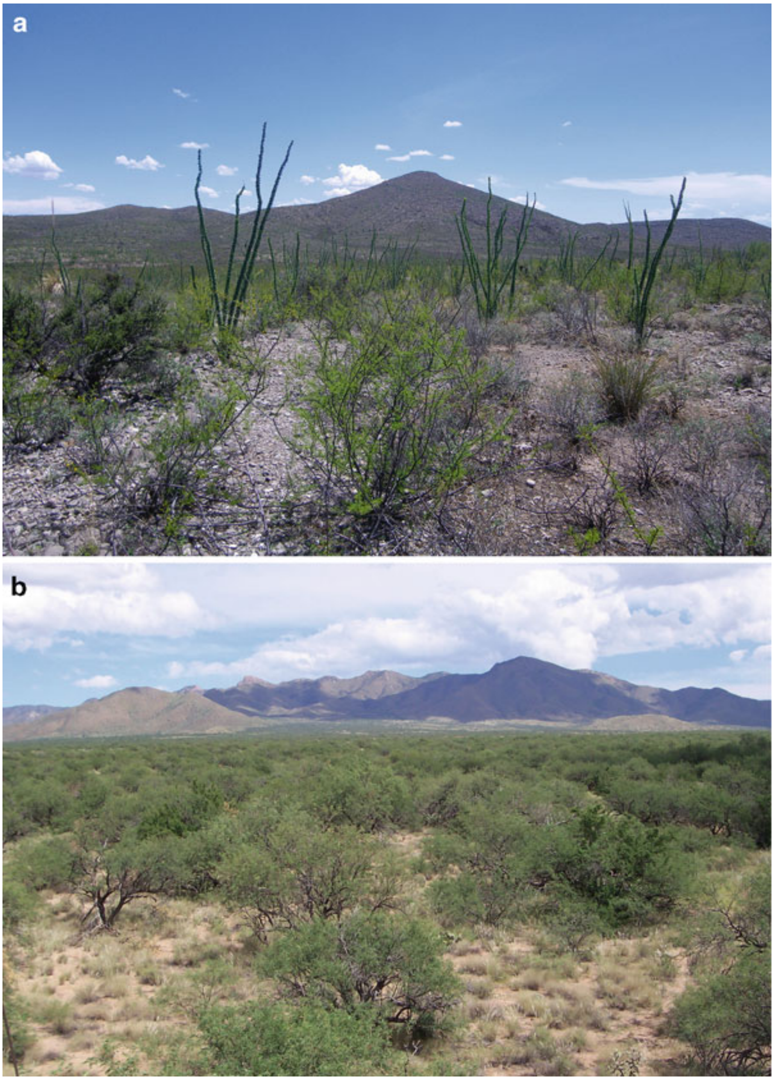

Fig. 15.2 Drylands in Arizona (USA): (**a**) Near Tombstone (by E. P. Sánchez-Cañete) and (**b**) Santa Rita Mesquite Savanna (by Russel L. Scott)

15.5 Main Natural Processes Involved in Carbon Sequestration and Loss

Estimates of C pools mentioned above are the result of several processes which contribute to the net ecosystem CO_2 exchange (NEE) with the atmosphere in drylands.

Fig. 15.3 Tabernas desert located in Almería, Southeast Spain (by C. Oyonarte)

15.5.1 Biological Processes

The SOC stored in soils and that in the biomass are mainly due to the balance between photosynthesis (net CO_2 uptake) and respiration (net CO_2 release via decomposition, degradation and diffusion processes). Such biological processes are mainly responsible for annual NEEs in most ecosystems (forest, wetlands, cropland, etc.). Thus, the FLUXNET community (Baldocchi et al. 2001) interprets CO_2 fluxes measured using micrometeorological techniques (Dabberdt et al. 1993) as a biological flux neglecting non-biological processes (Falge et al. 2002; Reichstein et al. 2005; Stoy et al. 2006; Valentini et al. 2000). However, many CO_2 flux measurements over drylands indicate contributions of abiotic processes to the NEE (Emmerich 2003; Ferlan et al. 2011; Hastings et al. 2005; Inglima et al. 2009; Mielnick and Dugas 2000; Were et al. 2010; Wohlfahrt et al. 2008; Xie et al. 2008). These processes can dominate the flux during the dry season (Kowalski et al. 2008) with annual contributions >50% depending mainly on meteorological conditions (Serrano-Ortiz et al. 2009).

15.5.2 Weathering Processes

Some soils are derived from carboniferous (calcareous) rocks and/or include additional carbonates (secondary carbonates or caliche) as a result of weathering and precipitation processes (Eq. 15.1).

$$H_2O_{(aq)} + CO_{2(g)} + CaCO_{3(s)} \Leftrightarrow 2HCO_3^- + Ca^{2+}_{(aq)} \tag{15.1}$$

In terms of NEE, Eq. 15.1 specifies that for each molecule of $CaCO_3$ dissolved a molecule of atmospheric CO_2 is consumed. During precipitation, one molecule of CO_2 is released for every molecule of $CaCO_3$ deposited to the surface. Thus, either water lost by evapotranspiration or additional sources of Ca^{2+} would enrich aqueous concentrations and enhance the deposition of $CaCO_3$ from the aqueous solution, and release CO_2 to the atmosphere (Eq. 15.1 to the left).

Globally, and over long time scales, weathering processes are balanced with respect to CO_2 (Berner 2003; Lasaga et al. 1994). However, at annual and seasonal scales the predominance of dissolution or precipitation processes may be relevant to local NEEs (Serrano-Ortiz et al. 2010) and contribute to the observed CO_2 fluxes (Emmerich 2003; Mielnick et al. 2005). In addition, although precipitation processes imply CO_2 release at short time scales, caliche formation can be considered a net atmospheric CO_2 sink over long timescales: every two molecules of bicarbonate, previously formed during the growing season by dissolution of two molecules of CO_2, react with one molecule of additional Ca^{2+} to form one molecule of calcite and release one molecule of CO_2. The formation of this secondary C form depends on land use and soil/crop management systems (Lal 2004). Addition of biomass, whose decomposition increases the partial pressure of CO_2 in the soil, together with irrigation, increases SIC in agricultural soils (Entry et al. 2004; Lal 2004). While its contribution to the total annual atmospheric CO_2 sink may be less than 10% (Eswaran et al. 2000; Gombert 2002; Liu and Zhao 2000; Mermut et al. 2000), it is unclear how SIC responds to rainfall and temperature changes predicted under the climate change scenarios (Rasmussen 2006).

15.5.3 Bio-sequestration

Fungi, lichens, and cyanobacteria play a prominent part in calcite dissolution and precipitation by biomineralization processes (Verrecchia et al. 1999). In addition, one inert form of organic C can be bio-sequestered within plants and accumulates in soil after the decomposition of that vegetation (phytolith-occluded C) (Parr and Sullivan 2005). Although phytoliths are highly resistant to oxidation and very stable in soil, they can be involved in SOC dynamics in response to land use and management change (Allen et al. 2010). There are many studies regarding the ability of different microorganisms to precipitate carbonate in drylands (Delgado et al. 2008; Li et al. 2011; Parraga et al. 2004; Rivadeneyra et al. 1997). However, little is known about its contribution in the soil C sequestration.

15.5.4 Ventilation

Drylands over carbonate rocks with cracks, pores and cavities, together with soils with deep vadose-zones, show a high capacity to store CO_2 belowground. Since such CO_2 storage may represent as much as 60% of the annual atmospheric sink (Serrano-Ortiz et al. 2011), the subsurface can be considered a temporal depot for CO_2 coming from different processes (mainly weathering and respiration) (Serrano-Ortiz et al. 2010). In addition to diffusion processes, such soils have the potential to emit the stored CO_2 via ventilation (Sanchez-Cañete et al. 2011) and contribute to ecosystem CO_2 exchange observed. Ventilation is a transport process due to net movements of air in and out of an enclosed space. The behaviour of ventilation processes, in caves for example, is controlled by the degree of connection between the cavities and the aboveground system (Cuezva et al. 2011) and, thus, such processes have only been detected when the soil is dry (Cuezva et al. 2011; Sanchez-Cañete et al. 2011; Serrano-Ortiz et al. 2009; Were et al. 2010). The main meteorological drivers controlling soil CO_2 ventilation due to pressure pumping are wind speed and turbulence (Jassal et al. 2005; Lewicki et al. 2010; Subke et al. 2005; Takle et al. 2004). Therefore, the non-negligible role of subsurface as a temporal depot of CO_2, along with seasonal ventilation can contribute to the annual net ecosystem C balance (Serrano-Ortiz et al. 2010).

15.5.5 Erosion

Erosion is a natural process that occurs when a liquid (air or water) moves into and/or across a soil surface with subsequent transport of the detached particles to another location (Flanagan 2006) reducing the amount of SOC in the eroded soil. This process is more important in drylands with lower percentages of vegetation cover, and human activities associated with agricultural practices usually enhancing erosion. There is a lack of agreement whether water erosion induces net release of C to the atmosphere (Jacinthe and Lal 2001; Lal et al. 2004) or net C sequestration (Boix-Fayos et al. 2009; Harden et al. 1999; Van Oost et al. 2007), with estimates ranging from a source of 1 Pg C year^{-1} to a sink of the same magnitude. The SOC displaced in terrestrial ecosystems and mineralization during water transport can lead to CO_2 emissions, with global estimates of 56–168 g C m^{-2} year^{-1} and 6–52 g C m^{-2} year^{-1}, respectively (Jacinthe and Lal 2006). On the other hand, SOC exported from the eroded areas is replaced by additional C derived from the atmosphere providing a sink of atmospheric CO_2 (Van Oost et al. 2007). A global C sink of 0.12 Pg C year^{-1} is estimated to result from erosion in the world's agricultural lands (Van Oost et al. 2007).

15.5.6 Photodegradation

The direct breakdown of organic matter by ultraviolet (UV) light (photodegradation) contributes to litter mass loss (emission of CO_2) in water-limited ecosystems receiving intense sunlight (Austin and Vivanco 2006; Rutledge et al. 2010).

Therefore, future climate changes in radiation due to decreased cloudiness or increased stratospheric ozone depletion may have an effect on the C balance in such ecosystems (Austin and Vivanco 2006). Although photodegradation contribute to organic matter decomposition via microbial facilitation, the direct breakdown of organic matter to CO_2 can occur in the absence of microbial activity (Brandt et al. 2009). Thus, organic matter decomposition is not restricted to periods of high moisture availability as is plant production (Gallo et al. 2009). During midday in summer, the CO_2 efflux due to photodegradation contributes around 90% of the total half-hourly CO_2 flux from an arid grassland (Rutledge et al. 2010). However, the relevance of photodegradation and its contribution to the total CO_2 losses at ecosystems scales is still unknown. While Rutledge et al. (2010) estimates a C loss of 16 g m^{-2} for the dry season in an arid grassland located in the lower foothills of the Sierra Nevada (USA) using chambers and *eddy covariance* measurements, an extrapolated laboratory study to field conditions reveal an annual C emission of 4 g m^{-2} due to photodegradation for a desert grassland located in New Mexico (Brandt et al. 2009). Further studies are needed to increase the understanding, importance and drivers of photodegradation.

15.6 Carbon Sink Capacity at Ecosystem Level

The processes mentioned above (weathering, ventilation and/or erosion processes) act together in drylands and contribute to the measured annual net C exchange (Table 15.2). For deserts located in southwest of the U.S.A. and Baja California, published studies have determined that the most important driver controlling CO_2 flux is the not the amount of rainfall but mostly its timing (Hastings et al. 2005; Mielnick et al. 2005; Wohlfahrt et al. 2008). However, the published data on annual net C exchanges do not support this hypothesis. While two desert shrubland located in the Mojave Desert and Baja California with similar annual precipitation act as annual net C sinks of 106 ± 70 and 52 g C m^{-2} year^{-1} respectively, the Chihuahuan Desert site emits ~145 g C m^{-2} annually. Grasslands located in North America (New Mexico, Arizona and California) and Europe (Southeast Portugal and Southwest of Spain) are C sources ranging from 141 (source) to −190 (sink) g C m^{-2} year^{-1} depending mostly on the total amount of rainfall (Aires et al. 2008; Anderson-Teixeira et al. 2011; Emmerich 2003; Ma et al. 2007; Scott et al. 2006) and also wind speed for the particular site located in Southwest of Spain (Rey et al. 2012). While ecosystem C sink capacity in grasslands located in Northern Asia also depend on optimal temperature in summer (10–20°C) (Kato et al. 2006; Wang et al. 2008). For savannas, Scott et al. (2009) measured an annual net C releases ranging from 14 to 95 g C m^{-2} year^{-1} in a semiarid savanna in southern Arizona, while Ma et al. (2007) measured an annual net C uptake ranging from 56 to 155 g C m^{-2} year^{-1} in a savanna site located in California with higher annual precipitation and lower temperature. Finally, shrublands in drylands act mostly as small sinks for atmospheric CO_2 (uptake from 2 to 75 g C m^{-2} year^{-1}) (Anderson-Teixeira et al. 2011; Luo et al. 2007;

Table 15.2 Annual net ecosystem C exchange (g C m^{-2}) measured mostly using the eddy covariance technique, together with annual temperature and rainfall

Vegetation	Location	Experimental Site	Reference	Mean annual temperature (°C)	Mean annual precipitation (mm)	Net ecosystem C exchange (g C m^{-2} $year^{-1}$)
Desert and bare soil	Mojave Desert (USA)	Desert on the Nevada Test Site, 120 km northwest of Las Vegas	Wohlfahrt et al. (2008)	20	210	−106 ± 70
	Chihuahuan Desert (USA)	About 40 km northeast of Las Cruces, New Mexico	Mielnick et al. (2005)	–	272	145[a]
	Baja California (Mexico)	15 km west of the city of La Paz and 1.5 km from the Bay of La Paz (CIBNOR)	Hastings et al. (2005)	24[b]	147	−39
					197	−52
Grassland	Southwest Spain	*"Cabo de Gata Natural Park" Almería (Andalucía)*	Rey et al. (2012)	17	210	66
					251	144
					294	92
	Southeast Portugal	*Monte do Tojal*, Évora in Southern Portugal	Aires et al. (2008)	14.7	364	49
					751	−190
	New Mexico (USA)	Sevilleta LTER in Central New Mexico	Anderson-Teixeira et al. (2011)	13	244	30
	Arizona (USA)	The Kendall grassland Agricultural Research Service Walnut Gulch Experimental Watershed	Scott et al. (2010)	14.5	313	−69
					312	−98
					274	−55
					246	−47
					162	21
			Emmerich (2003)	17	356	126[a]
		Foodplain terraces along the San Pedro River	Scott et al. (2006)	17	234	−63
	California (USA)	Foothills of the *Sierra Nevada*	Ma et al. (2007)	17 ± 1	562 ± 193	(−88)–141
	Qinghai-Tibetan Plateau (China)	Alpine meadow	Kato et al. (2006)	−0.65	561[c]	−79
				−0.91		−92
				−1.53		−173
	Northern China	Inner Mongolia Autonomous region	Wang et al. (2008)	2.5	297	10
				1.3	174	30
				1.7	215	−15
	Central Mongolia	Hentiy province of Mongolia	Li et al. (2005)	1.2	196	−41

Savanna	Arizona (USA)	The Santa Rita mesquite savanna site located on the Santa Rita Experimental Range (SRER)	Scott et al. (2009)	19	285	60
				20	335	14
				20	289	95
				19	330	30
	California (USA)	Foothills of the *Sierra Nevada*	Ma et al. (2007)	17 ± 1	562 ± 193	(−155) – (−56)
Shrubland	Southwest Spain	Mediterranean plateau, 25 km from the coast	Serrano-Ortiz et al. (2009)	12	475	−2 ± 23
	New Mexico (USA)	Sevilleta LTER in Central New Mexico	Anderson-Teixeira et al. (2011)	14	244	−30
	San Diego (USA)	Southern California, 75 km east of Pacific Ocean	Luo et al. (2007)	15	349	−52
	Arizona(USA)	Foodplain terraces along the San Pedro River	Scott et al. (2006)	17	234	−212
		Lucky Hills, Agricultural Research Service Walnut Gulch Experimental Watershed	Emmerich (2003)	17	256	144[a]
	Qinghai-Tibetan Plateau (China)	Alpine *Potentilla fruticosa* at the Haibei Research Station	Zhao et al. (2006)	2.3	542	−59
				2.2	493	−75

Negative sign (−) for net ecosystem exchange indicates a C sink
[a]Data not measured by the eddy covariance technique
[b]Estimated data using information of Fig. 15.1 of the cited reference
[c]Annual average precipitation for 1981–2000

Serrano-Ortiz et al. 2009; Zhao et al. 2006), with some exceptions. In riparian areas where shrubs and woody plants have the capacity to exploit water resources by growing deep roots (Domingo et al. 1999), annual NEE can be higher than 200 g C m^{-2} $year^{-1}$ (Scott et al. 2006). On the contrary, the annual C loss estimation for the Lucky Hills site located in Arizona is 144 g C m^{-2} $year^{-1}$. The source of this C appears to be from the large SIC pool in these soils (Emmerich 2003). In summary, according to published studies, the C sink capacity at ecosystem level in drylands is highly variable depending on the ecosystem type, SIC pool and mostly on rainfall timing and temperature during the growing season.

15.7 Management Practices

Human activities are directly or indirectly responsible for dryland degradation, but have also the capacity to alter natural processes involved in C sequestration with the potential to ameliorate poverty-provoking desertification highly linked to poverty (Glantz 1994; Mouat and Lancaster 2006). Cultivated lands are about 30% of drylands contributing 20% of the total SOC pool. Thus, better management practices in cultivated lands (mainly croplands) could improve the role of drylands in the storage of SOC. Sequestration of C in croplands can be improved if NT or minimum-tillage practices are applied instead of the conventional systems (Table 15.1) (López-Fando and Pardo 2009, 2011; Sombrero and de Benito 2010). In addition, crop residues left on the soil surface instead of being removed or incorporated into the soil may increase the SOC storage by more than a 30% (Álvaro-Fuentes and Paustian 2011). Also, continuous cultivation instead of leaving land fallow could increase SOC pools by more than twice (Álvaro-Fuentes and Paustian 2011). To prevent soil erosion, croplands with slopes greater than 15% should be converted to grasslands. In this context, land use conversion from cropland to shrubland or wild grassland would be better for SOC sequestration than tree plantation in semi-arid lands (Chen et al. 2007). For grasslands, SOC storage decreases substantially by grassland degradation due to long-term heavy grazing. At least two decades of grazing prohibition would be appropriate for restoring grasslands from degraded to undisturbed natural SOC conditions (He et al. 2008). However, such direct interventions and control policies should be based on reliable ecological and economic arguments (Puigdefábregas 1998).

15.8 Conclusions

Analyses of the published literature concerning the carbon cycle over drylands support the following conclusions:

1. Although numerous studies have been published related to SOC pools and C in biomass, a conclusive global analysis using models is needed to provide credible estimates of C pool in drylands.

2. However, an approximate value of 230 Pg of SOC in drylands is widely accepted. This value is reduced by 15% taking into account the percentage of cultivated lands.
3. Also, an approximate value of 65 Pg of total C content in biomass is widely reported. This value can also be modified significantly taking into account the anthropogenic practices.
4. Implications for the drylands C sink capacity of geochemical processes (relevant for SIC) or ventilation and erosion are poorly understood.
5. Thus, the differences in the measured net C exchange with the atmosphere, ranging from −106 to 145 $g\,C\,m^{-2}$ for deserts and from −190 to 140 $g\,C\,m^{-2}$ for grasslands are not easily explained.
6. Less drastic anthropogenic land use such as NT or minimum-tillage agricultural practices, leaving residues on the soil surface, or reducing fallow croplands and temporary grazing exclusions in grassland may improve SOC sequestration in drylands. Ideally, NT management may potentially increase the SOC by 20%, while non-grazing in grassland could increase SOC storage by about 45%.

Acknowledgements PSO is a Juan de la Cierva fellow, funded by the Spanish Ministry of Science and Innovation. English revision was provided by Professor Kowalski. Comments from A. S. Kowalski, Ana Were, Francisco Domingo, Gabriel Delgado and Russel L. Scott improved this paper.

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