Equation for compaction of paleosols due to burial

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ABSTRACT

Existing empirically derived compaction curves for marine sediments fail to estimate the compaction of paleosols, flood-plain sediments, and peats as indicated by deformation of clastic dikes, dinosaur footprints, and sandstone paleochannels. Yet compaction estimates are vital to reconstructing paleoclimate and geochemical profiles of paleosols, to modeling sediment accumulation rates of nonmarine rocks, and to tectonic flexural modeling of molasse facies. This paper presents a compaction equation and physically derived constants that make geologically realistic estimates of compaction of paleosols and other nonmarine sediments. A protocol for the application of the equation is suggested that would allow the following equation to solve for burial compaction (C as a fraction of original thickness) given depth of burial (D in km) as: $C = -S_i/[(F_0/\ e^{Dk}) - 1]$. This equation can be applied to nonmarine sedimentary rocks or paleosols given appropriate empirical data on the physical constants in the equation, such as initial solidity (S_i), initial porosity (F_0), and the corresponding curve-fitting constant (k). Data on physical constants useful for this equation are compiled here for a range of paleosol and nonmarine sediment types.

Keywords: compaction, paleosols, burial, peat, flood plains.

INTRODUCTION

Compaction due to burial can be estimated from equations fitted to empirically derived data for most types of marine sedimentary deposits (Baldwin and Butler, 1985). Similar empirical relationships have not been derived for paleosols, yet are needed in order to reconstruct original geochemical profiles of paleosols (Retallack, 1986), to estimate paleoprecipitation from the depth to the calcic horizon (Retallack, 1994, 1997; Caudill et al., 1997), and for tectonic-flexural and sequence stratigraphic modeling of nonmarine sedimentation (e.g., Driese et al., 1994). In this paper we review shortcomings of currently used compaction equations and present a general equation that can be used for paleosols and nonmarine sediments.

Retallack (1993) used an equation derived by Baldwin and Butler (1985) from studies of marine sandstone compaction by Sclater and Christie (1980). However, measurements of compaction of clastic dikes in paleosols of known burial depth by Caudill et al. (1997) showed that the sandstone curve of Baldwin and Butler (1985) as applied by Retallack (1993, 1994, 1997) yielded unrealistically high compaction of paleosols at depth. Caudill et al. (1997) offered an alternative equation that improves compaction predictions for Vertisols buried to depths of 2–10 km. Unfortunately, this equation makes the geologically unrealistic prediction that Vertisols would expand if buried <1.82 km. The alternative constants chosen by Caudill et al. (1997) are related to intrinsic physical properties of soil and sediment such as grain density and bulk density. Nadon (1993) and Nadon and Issler (1997) also showed, from compaction of dinosaur footprints and paleochannels, that existing compaction curves based on marine rocks give unrealistically high compaction. They also showed that physical properties vary considerably for different soils and sediments from the values chosen by either Retallack (1993) or Caudill et al. (1997).

Any attempt to account for compaction of paleosols or floodplain sediments either for paleopedological purposes or for tectonic-flexural and sequence stratigraphic modeling should apply to all types of paleosols and at all burial depths. Retallack's (1993) original equation is reasonable for paleosols such as Inceptisols or sediments such as siltstones with bulk densities around 1.3 g·cm^{-3} (which translates to a S_i value of 0.5, see following), but given the range of bulk densities in Table 1, it is insufficient for all types of paleosols and floodplain sediments. Caudill et al.'s (1997) equation is only intended for use with Vertisols buried 2–10 km. Unfortunately, it predicts expanding paleosols at burial depths <1.82 km, when it is clear from evidence such as contorted clastic dikes that these too have been compacted. Nonetheless, for deeply buried Vertisols, Caudill et al.'s (1997) equation predicts compaction values that approximate empirical compaction observations. In light of the shortcomings in the applicability of Retallack's (1993, 1994) and Caudill et al.'s (1997) equations, we have derived the equation in general terms and compiled a database of constants for a variety of paleosol and floodplain sediment types.

PALEOSOL COMPACTION EQUATIONS

Sclater and Christie (1980) presented a curve fitted to the relationship between burial porosity and measured surface porosity as an exponential function. Baldwin and Butler (1985) used their data and equation to recast the Sclater-Christie exponential equation for sandstone in logarithmic form:

$$D = 3.7\ln[0.49/(1 - S_b)], \tag{1}$$

where D is burial depth (in km) and S_b is burial solidity (fractional complement of porosity). Retallack (1991) suggested that this sandstone equation could be applied to paleosols more accurately than the shale equation derived from the Sclater-Christie equation, because the marine shales on which it was based have a much higher initial water content than paleosols. The ped structure of soils, consisting of soil clods surrounded by cracks, was envisaged as physically more like a sandstone or grain-supported conglomerate than a mudstone. Retallack (1993) also assumed that compaction is governed by the relationship:

$$C = S_i S_b^{-1}, \tag{2}$$

TABLE 1. CONSTANTS FOR THE APPLICATION OF THE COMPACTION EQUATION

Substrate	Density	Range	σ	S_i	F_0	k	Source
Marine							
Shale	1.07	—	—	0.37	0.63	0.51	1
Sand	1.35	—	—	0.51	0.49	0.27	1
Chalk	0.81	—	—	0.30	0.70	0.71	1
Shaley Sand	1.18	—	—	0.44	0.56	0.39	1
Soil Types							
Alfisol (n=46)	1.68	1.33–1.97	0.16	0.65	0.35	0.15	2
Andisol (n=26)	0.79	0.44–1.50	0.27	0.30	0.70	0.71	3, 4, 5
Aridisol (n=24)	1.60	1.39–1.76	0.09	0.62	0.38	0.17	2
Entisol (n=3)	1.61	1.60–1.64	–	0.62	0.38	0.17	2
Histosol (n=13)	0.07	0.04–0.10	0.02	0.06	0.94	2.09	6, 7
Inceptisol (n=41)	1.32	0.65–1.92	0.39	0.51	0.49	0.27	2
Mollisol (n=50)	1.42	0.85–1.91	0.25	0.55	0.45	0.23	2
Oxisol (n=31)	1.30	0.96–1.46	0.11	0.50	0.50	0.29	2, 8
Spodosol (n=20)	0.97	0.30–1.87	0.47	0.37	0.63	0.52	2
Ultisol (n=38)	1.50	0.97–1.84	0.27	0.58	0.42	0.20	2
Vertisol (n=380)	1.80	1.55–2.06	0.16	0.69	0.31	0.12	9
Modern Floodplain							
Inorganic Silts and Clays							
Mean (Liquid limit <50)*	—	—	—	0.635	0.365	0.16	10
Mean (Liquid limit >50)†	—	—	—	0.511	0.489	0.27	10
Sands							
Mean§	—	—	—	0.692	0.308	0.12	10

Note: Units of density and range are g·cm^{-3}. Units on k are x10^{-5} cm^{-1}. Sources for the data contained in the paper are: 1. Sclater and Christie (1980); 2. Soil Survey Staff (1975); 3. Shoji et al. (1993); 4. Shoji et al. (1988); 5. Bielders et al. (1990); 6. Subekty et al. (1993); 7. Weller (1959); 8. Muchena and Sombroek (1981); 9. Eswaran et al. (1989); 10. Nadon and Issler (1997).

*Silts (n = 61) and clays (n = 261); porosity ranges from 35.72% to 37.34%.

†Silts (n = 9) and clays (n = 61); porosity ranges from 46.96% to 50.80%.

§Clean (graded; n = 20), clean (poorly graded; n = 62), silty sands (n = 153), and clayey sands (n = 88); porosity ranges from 30.10% to 31.59%.

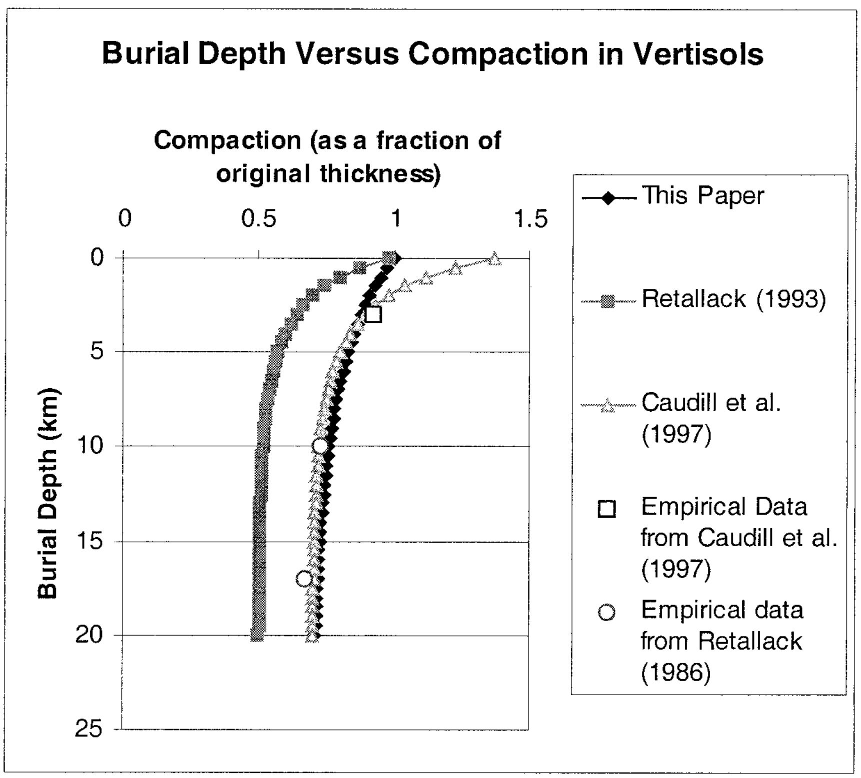

Figure 1. Plot of equations of Retallack (1993), Caudill et al. (1997), and this paper relating burial depth to compaction as percentage of original thickness. Note that while Retallack (1994) curve rapidly overcompacts sediments, Caudill et al. (1997) curve suggests expanded sediments buried at depths of < ~2 km. New curve and equation presented in this paper neither overcompact nor expand sediments buried at any depth.

where C is compaction (as a fraction of original thickness) and S_i is initial solidity. From this, he recast the equation in order to estimate compaction from geologically based estimates of the depth of burial

$$C = -S_i/[0.49/e^{(D/3.7)} - 1]. \tag{3}$$

Retallack (1993) suggested that an initial solidity (the complement of initial porosity) of 0.5 was reasonable assuming an average dry bulk density (ρ_d) of 1.3 g·cm^{-3} for soils and using 2.6 g·cm^{-3} (i.e., montmorillonite) for the solid grain bulk density (ρ_s) in the following relationship:

$$S_i = \rho_d \rho_s^{-1}. \tag{4}$$

Caudill et al. (1997) disputed the initial solidity value of 0.5 and suggested instead a value of 0.7 because Eswaran et al. (1989) calculated a dry bulk density for Vertisols of 1.8 g·cm^{-3}. They supported this change by showing that it predicts well the burial of several deeply buried Vertisols. Caudill et al. (1997) preferred a compaction equation of:

$$C = -0.7/[0.49/e^{(D/3.7)} - 1]. \tag{5}$$

Caudill et al.'s (1997) equation is a better approximation for deeply buried Vertisols, but it does not work for shallowly buried Vertisols, and is not intended for use with other paleosol types. At the surface (D = 0), we would expect a compaction equation to show no fractional compaction (C = 1). Caudill et al.'s (1997) equation yields a geologically unrealistic compaction value of 1.37 for unburied paleosols. Caudill et al.'s (1997) equation does not give a C value of one until a depth of 1.82 km (Fig. 1). Caudill et al. (1997) took compaction values of greater than one to represent uncompacted Vertisols; they postulated that the high initial solidity of Vertisols would preclude much compaction until they had been buried to a depth of more than 1.5 km. Compaction equations are more useful if applicable to any depth range and to more than just one paleosol type.

COMPACTION EQUATION

In order to develop a widely applicable compaction equation, consider the original equation of Sclater and Christie (1980) for normally pressured sections:

$$F = F_0 e^{-kD}, \tag{6}$$

where F is burial porosity (as a fraction), F_0 is initial porosity (as a fraction), D is depth (km), and k is an empirically derived constant. Equation 6 can be recast to a logarithmic equation:

$$\ln(F_0/F) = kD. \quad (7)$$

Solidity is the complement of porosity, given by these expressions:

$$S_i = 1 - F_0. \quad (8)$$

$$S_b = 1 - F. \quad (9)$$

Combining equation 2 with equation 9 and putting the result in equation 7 we get:

$$\ln [F_0/(1 - S_i/C)] = kD. \quad (10)$$

Dividing both sides of the equation by k, we get the general form of Baldwin and Butler's (1985) equation 1. Following Retallack (1993), we can estimate compaction using the following generalized equation:

$$C = -S_i/[(F_0/ e^{Dk}) - 1]. \quad (11)$$

Applications of equation 11 require careful consideration of the values for constants for a given soil or sediment (Table 1). Using the relationships for marine sediments outlined by Sclater and Christie (1980), the constant k can be related to initial porosity (F_0) by the following equation:

$$k = 0.03e^{4.52Fo}. \quad (12)$$

The constant k is a curve-fitting constant in the original Sclater-Christie equation, and as such, is based on the empirical relationship between initial porosity and burial porosity (see Appendix A of Sclater and Christie, 1980).

APPLICATIONS OF THE COMPACTION EQUATION

Using the dry bulk density of Vertisols determined by Eswaran et al. (1989) and equation 4 we get an initial solidity of 0.69 (the unrounded value of Caudill et al., 1997). Given that solidity is the complement of porosity, we use equation 8 to find an F_0 value of 0.31. Inserting 0.31 into equation 12, we get a k value of 0.12 for Vertisols. These values in the h;ompaction equation 11 give an equation that can be used to calculate the compaction of Vertisols as follows:

$$C = -0.69/[(0.31/e^{0.12D}) - 1]. \quad (13)$$

Reconsidering the Vertisol studied by Caudill et al. (1997), who assumed a burial depth of 3 km, we get a compaction value of 88% ($C = 0.88$), which compares favorably with observed deformation of its clastic dikes of 92%–93% (Fig. 1). Furthermore, when this equation is applied to Vertisols buried 10–17 km, it yields compaction values of 72%–76% (at depths of 17 and 10 km, respectively), which compares fairly well with Retallack's (1986) estimates of 67%–73% from compaction of clastic dikes. Equally important, equation 13 shows no compaction ($C = 1$) for unburied paleosols and/or sediments.

A second example is the compaction of peat to form coal. As Nadon (1998) pointed out, the compaction of peat has frequently been overestimated. By examining paleochannel geometry and dinosaur footprints, Nadon (1998) showed that peat is not as easily compactible as widely thought, and that most of the compaction of peat occurs during accumulation and at very shallow burial depths. Using bulk density data of Subekty et al. (1993) and a typical solid bulk density of 1.11 g·cm^{-3} (Weller, 1959), we get a S_i value of 0.06. The corresponding F_0 value is 0.94; using equation 12, the corresponding k value is 2.09. The final compaction equation for peat is:

$$C = -0.06/[(0.94/e^{2.09D}) - 1]. \quad (14)$$

Nadon (1998) described a peat with a compaction ratio of 1.15 that was buried just 3 m. Solving equation 14 for compaction at a depth of 0.003 km, a value of 0.9125 is obtained. Inverting this value we find a compaction ratio of 1.09, which is fairly close to the value of 1.15 given by Nadon (1998) (for reference, that value represents a compaction of about 87% of the original thickness).

In addition to the worked examples presented here, a better understanding of the compaction response of paleosols and nonmarine sediments to burial or loading, and of the physical properties of soils (as in Table 1), has a number of other applications. Flexural modeling (e.g., Flemings and Jordan, 1990) of paleosol-dominated sections such as the Siwalik Group in the Himalayan foothills of Pakistan and India often involves overestimation of initial sediment bulk density (often assumed to be 2.3 g·cm^{-3}, while the highest mean value in Table 1 is 1.8 g·cm^{-3}), which leads to overestimates of both flexure and the creation of accommodation space. Attempts to constrain sediment accumulation rates could also be improved. For example, Kraus and Bown (1993) used the data of Baldwin and Butler (1985) and assumed that the Willwood Formation of Wyoming was 25% sandstone and 75% mudstone when estimating the compaction of the section. But the mudstone is mostly paleosols of a variety of pedotypes (Retallack, 1998), and their different physical properties significantly affect the sediment accumulation rate model. There are also economic applications because paleosols have been used to determine oil reservoir dimensions (e.g. Ye, 1995) and to define sequence stratigraphic boundaries (e.g., McCarthy and Plint, 1998; Driese et al., 1994). In most such cases paleosols have been treated as siltstones or sandstones, and improved estimates of compaction should consider the unique characteristics of paleosols and their range of physical variability.

RECOMMENDED PROCEDURE

A compaction equation derived from the Sclater-Christie curve fit can be applied to paleosols and nonmarine sediments, provided it is constrained by physically realistic constants. The following protocol is suggested for the application of the compaction equation.

1. Select an initial solidity value (S_i) for comparable modern soils or sediments (Table 1) and find the corresponding porosity (F_0) by considering both the dry bulk density (ρ_d) and the solid bulk density (ρ_s) of the substrate in question.

2. Use the initial solidity value to solve for the appropriate constant k using equation 12.

3. Using physically reasonable constants in the compaction equation, solve for the compaction at the given depth of burial indicated by local stratigraphic or thermal maturity evidence.

CONCLUSIONS

By rederiving the exponential porosity equation of Sclater and Christie (1980) in general terms, it is possible to find a compaction equation that can yield results that are close to compaction estimates from empirical data, such as observed compaction of clastic dikes (Fig. 1). Careful consideration of the values for initial solidity (porosity) is essential to any attempt to characterize the compaction of paleosols (Table 1). Such reconstruction of original soil thickness is useful for a variety of interpretive studies of paleosols (Retallack, 1986, 1997; Caudill et al., 1997). While we have emphasized the application of the compaction equation to paleosols, it can also be applied to nonmarine sediments by using reasonable physical constants in the equation. Use of lithology-specific constants as advocated here may yield better estimates of compaction than the Sclater-Christie constants do for marine sediments as well, and may give better results in decompaction and back-stripping calculations that include nonmarine sedimentary rocks and paleosols.

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